Institute of Water and Environment

PhD Thesis

2003

CATHERINE PEACOCK

REEDBED HYDROLOGY AND WATER REQUIREMENTS

CRANFIELD (Silsoe Campus)

Supervisor: Dr. T.M. Hess

February 2003

Abstract

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Reedbed Hydrology and Water Requirements

Stodmarsh National Nature Reserve includes the largest reedbed in Southern England and is an important habitat for breeding waders and several rare bird species, including Bitterns. A succession of drought years in the 1990s brought the issue of the hydrology and water requirements of the wetland to the attention of managers and there is concern about future water supplies to the reserve. This study aims to calculate the amount of water required by the site in order to maintain optimum habitat conditions. The greatest area of uncertainty in the water balance is the evapotranspiration rate of the reedbeds and therefore a secondary aim is to increase understanding of this flux. Detailed hydrological measurements were carried out over two years to establish the water balance of the site. Evapotranspiration was measured using the Bowen ratio technique, accompanied by additional physiological and meteorological measurements. Results showed that evapotranspiration from reeds was generally less than reference evapotranspiration and that stomatal resistance was the most important factor controlling evapotranspiration rates. The hydrology of the site was modelled using a thirty year historical data series to quantify the return periods of flood and drought conditions of different severity. These were used to predict water resource requirements and availability and confidence limits were attached to the results. In 70% of years, summer deficits in the rainfall-evapotranspiration balance require the addition of water from the Lampen Stream. In 10% of these years, the entire summer discharge of the Lampen Stream would be insufficient to meet site water requirements and an additional source of water is required. Competition with other water users and limits on abstraction will increase the number of years an additional water source is required. In addition future climate change is likely to increase summer water requirements whilst decreasing resource availability.

Acknowledgements

I would like to gratefully acknowledge English Nature, as the main sponsors of this research. I thank them for allowing free access to the study site and in particular thank Paul Burnham, the previous site manager of Stodmarsh NNR for initiating the research and for his enthusiasm and advice in its early stages. Subsequently the current site manager, David Feast also gave helpful advice and practical guidance. Special thanks to him, Ian Hayes and Becky for their help installing the field equipment. Also to Chris Newbold and Theresa Bennett of English Nature who attended thesis committee meetings and gave useful input.

I also thank the Society of Wetland Scientists for the award of a Student Research Grant, which allowed continuation of fieldwork in the final year of the research.

Several organisations provided me with additional data. I am grateful to the British Atmospheric Data Centre which provided me with access to the Meteorological Office Land Surface Observation Stations Data, and particularly to Anabelle Menochet for her consistently prompt and efficient response to data enquiries. The Environment Agency provided me with rainfall data – thanks to Kevin Grimmett for sending data every month – and also information regarding abstraction licences on the Lampen Stream. I am grateful to Stephen Edge, the Geographic Information Manager for English Nature who supplied basemaps and levelling data for the site. Climate change scenarios were created from datasets supplied by UKCIP.

I am particularly grateful to my supervisor, Dr. Tim Hess for his encouragement and patience. Also to my thesis committee members, Prof. Peter Leeds-Harrison and Dr Paul Burgess whose support has been invaluable.

The help of Ian Littlewood of the Centre for Ecology and Hydrology, Wallingford in calibrating the parameters of the IHACRES model was invaluable and I thank him for the work he did and for allowing me to use the model.

I am grateful to the following people for their helpful advice in the use of the Bowen ratio techniques and for sharing their Bowen ratio experiences: Dave Campbell of Waikato University, New Zealand, Rick Allen of University of Idaho, USA, Mike Acreman of the Centre for Ecology and Hydrology, Wallingford, UK, Mark McGlinchey of the Swaziland Sugar Association and Kevin Gilman, hydrological consultant, UK.

I would like to acknowledge the large number of technical staff at Cranfield University, Silsoe who have gone out of their way to help me. In particular, Nigel Janes and Simon Stranks who would carry endless amounts of heavy equipment through mud and water, all for some chips at the seaside, and Roger Swatland who patiently taught me how to do things with wires.

Special thanks also to Merle Collins for assistance with the error analysis and for reading and correcting the manuscript and also to Jean Penney for the same task. I thank both my parents for their encouragement and support throughout my education and especially my husband Dave for his love over the last three years. Finally thanks are due to God, the creator of reedbeds and sustainer of all life. With Him anything is possible.

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Notation

Roman

a	thermal diffusivity
A	catchment area
a_I	polygon area
C	volume forcing constant
C_p	specific heat capacity of air (1.13 kJ kg^{-1} °C^{-1})
C_s	specific heat capacity of soil (MJ m^{-3}°C^{-1})
C_w	specific heat capacity of water (MJ m^{-3}°C^{-1})
D	vapour pressure deficit (kPa)
d	zero plane displacement height (m)
D_c	coefficient of determination
D_e	departure from mean for estimated residual mass curve.
D_o	departure from mean for observed residual mass curve
e	water vapour pressure (kPa)
E	Coefficient of efficiency
e_a	saturation water vapour pressure of air (kPa)
e_d	vapour pressure at dewpoint (kPa)
ET_c	crop evapotranspiration (mm)
ET_{ref}	reference evapotranspiration (mm)
f	temperature modulation factor (IHACRES model)
f	function (Jarvis model)
F_m	fraction soil mineral matter
F_o	fraction soil organic matter
F_w	fraction soil water
G	ground heat gain (W m^{-2})
g	acceleration due to gravity (9.81 m s^{-2})
g_a	atmospheric conductance (mm s^{-1})
g_c	canopy conductance (mm s^{-1})
G_I	ground water inflow
G_o	ground water outflow
G_{plate}	soil heat flux measured by flux plates (W m^{-2})
g_s	stomatal conductance (mm s^{-1})
g_{so}	maximum conductance in saturated air (mm s^{-1})
H	sensible heat loss (W m^{-2})
h	height of vegetation (m)
k	von Karman's constant (0.04)
k	timestep in IHACRES model
Kc	crop coefficient
K_H	turbulent transfer coefficient for heat (m^2 s^{-1})
K_V	turbulent transfer coefficient for water vapour (m^2 s^{-1})
L	level above Ordnance datum (m)
P	atmospheric pressure (kPa)
p	probability

P_n	precipitation (mm)
q_c	observed value
$\overline{q}_c$	mean observed value
q_e	estimated value
q_{est}	estimated value from regression line of q_c on q_e.
Q_k	streamflow output
Qp	photosynthetic photon flux density
R	specific gas constant (287 J kg^{-1} K^{-1})
r	rank
r_a	aerodynamic resistance (s m^{-1})
Ri	Richardson number
r_k	daily rainfall (mm)
r_l	mean stomatal resistance of a leaf (s m^{-1})
R_n	net radiation gain (W m^{-2})
r_s	surface resistance (s m^{-1})
Rs	solar radiation (W m^{-2})
r_{smax}	maximum conductance in saturated air (s m^{-1})
r_{smin}	minimum stomatal resistance in optimal conditions (s m^{-1})
Rso	short wave radiation for a clear sky day (W m^{-2})
s	slope of the SVP temperature curve (kPa°C^{-1})
S_I	surface inflow
s_k	catchment wetness index
S_o	surface outflow
T	air temperature (°C)
t	time (s)
T_{dew}	dewpoint temperature (°C)
T_k	temperature (K)
T_{ref}	reference temperature (°C)
T_s	soil temperature (°C)
T_w	water temperature (°C)
u	windspeed (m s^{-1})
$u*$	friction velocity (m s^{-1})
u_k	effective rainfall (mm)
V	storage volume (m^3)
x_f	the minimum fetch requirement for near neutral conditions (m)
z	height (m)
z^{-1}	backwards shift operator
z_h	height of air temperature and humidity measurements (m)
z_m	height of windspeed measurements (m)
z_o	roughness length governing transfer of heat and vapour (m)
z_{om}	roughness length governing momentum transfer (m)
z_s	depth of soil layer (m)
z_{sw}	depth of water (m)

Greek

β	Bowen ratio
δ	absolute error
γ	psychrometric constant ($kPa°C^{-1}$)
Δ	difference between measurements
ε	ratio of molecular weight of water to the molecular weight of dry air (0.622)
λ	latent heat of vaporisation of water ($MJ\ kg^{-1}$)
λE	latent heat loss ($W\ m^{-2}$)
ρ	air density ($kg\ m^{-3}$)
σ	Stefan-Boltzmann constant ($4.90\times10^{-9}\ MJm^{-2}K^{-4}d^{-1}$)
ϕ_M	stability function for momentum
τ_w	catchment drying time constant
ψ	leaf water potential
Ω	de- coupling factor

Abbreviations

BREB	Bowen Ratio Energy Balance
DEFRA	Department of the Environment, Food and Rural Affairs
ELADP	Ellipsoidal Leaf Distribution Parameter
ESA	Environmentally Sensitive Area
ET	Evapotranspiration
EU	European Union
GCM	Global Circulation Model
GPS	Global Positioning System
IHACRES	Identification of unit Hydrographs and Component flows from Rainfall Evaporation and Streamflow data
IPCC	Intergovernmental Panel on Climate Change
IRGA	Infra Red Gas Analyser
LAI	Leaf Area Index
MORECS	Meteorological Office Rainfall and Evaporation Calculation System
NCC	Nature Conservancy Council
NNR	National Nature Reserve
PPCM	Parametrically Parsimonious Conceptual Model
SAC	Special Area of Conservation
SPA	Special Protection Area
SSSI	Site of Special Scientific Interest
UK	United Kingdom
UKCIP	United Kingdom Climate Impacts Programme
WaSim	Water balance Simulation model
WFD	Water Framework Directive

CHAPTER 1 – Introduction

1.1 STODMARSH NATIONAL NATURE RESERVE

This study is focussed on Stodmarsh National Nature Reserve (NNR). Stodmarsh NNR is located in the south-eastern corner of the UK, 8 km north-east of Canterbury in Kent (51°8'N, 01°10'E) (Figure 1.01).

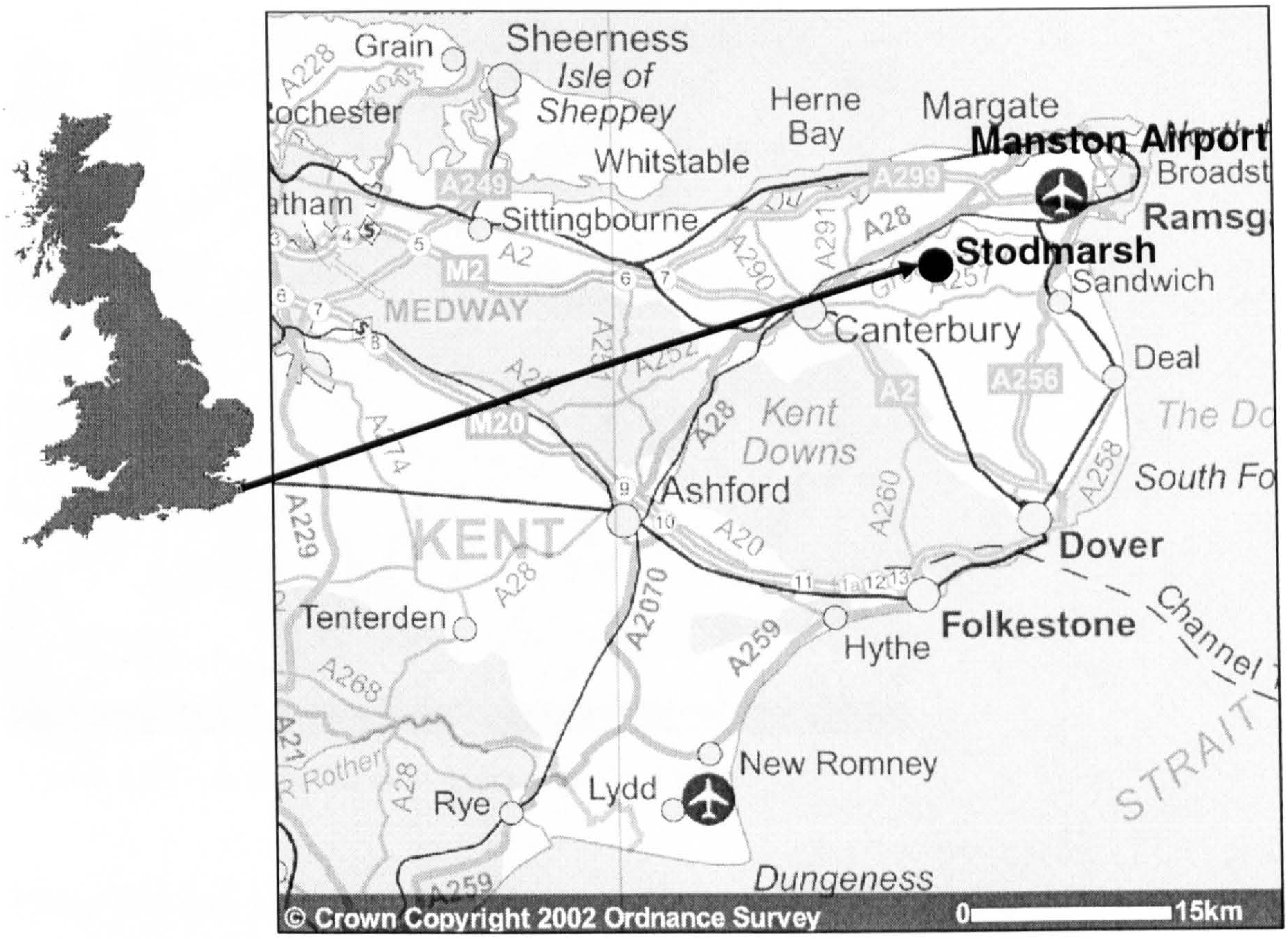


Figure 1.01 – Location map of Stodmarsh National Nature Reserve (●)

The site is owned and managed by English Nature. Stodmarsh National Nature Reserve, with an area of 320 ha, forms part of the larger Stodmarsh Site of Special Scientific Interest (SSSI) (604.4 ha) which is part of the flood plain of the River Great Stour. It has several habitat types – reedbeds, consisting mainly of *Phragmites australis,* which cover around two thirds of the reserve area, wet grazing meadows, covering around one third, and smaller areas of open water and alder carr. The reedbed is the rarest and most important habitat. The beds are the largest in the South of England and an area of national importance. The River Great Stour forms the boundary on the northern edge of the reserve, though it is not the water source of the wetland and bunds keep the two systems hydrologically distinct. The River Great Stour is shown in the more detailed map of Stodmarsh NNR in Figure 1.02. In 1995 English Nature, the site owners and managers, purchased an additional 79 ha of fields adjoining the NNR and reedbeds have been planted with the aim of encouraging Bitterns to breed as part of English Nature's Bittern Recovery Programme.

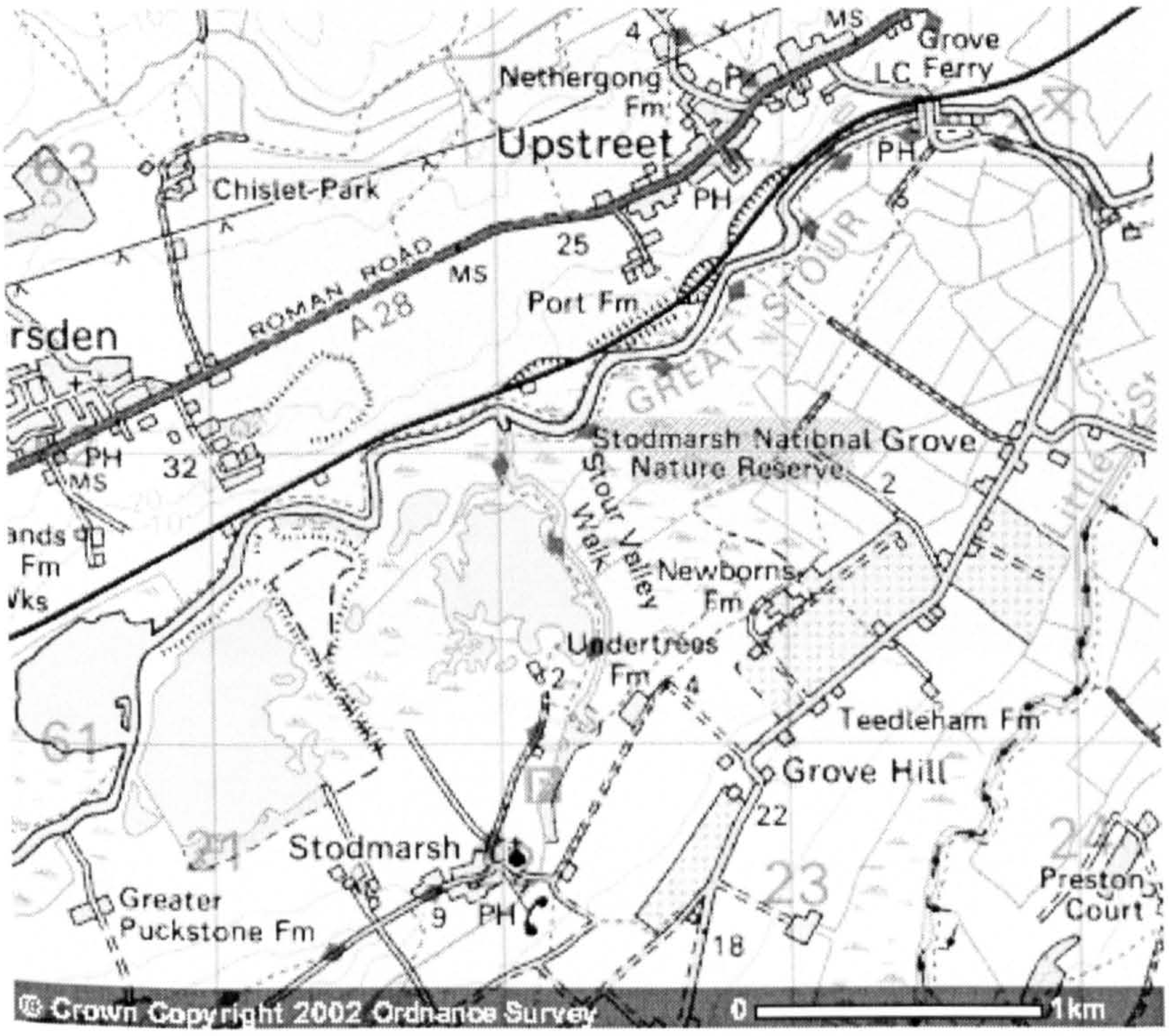


Figure 1.02 – A detailed map of Stodmarsh NNR

Plates 1.01 and 1.02 show the reedbeds at Stodmarsh NNR during the growing season and after senescence.

Plate 1.01 – Within the growing reedbed at Stodmarsh NNR

Plate 1.02 – Reeds at Stodmarsh NNR after senescence

1.1.1 Site history

Recorded history of the site began 1300 years ago when monks dug drainage ditches to bring floodwaters onto the meadows of the Stodmarsh area. The Lampen Wall – a clay bund running northwards through the site, was constructed in the eighteenth century as a flood defence. A colliery opened at the site at the start of the twentieth century and the land subsequently subsided over the underground workings causing lagoons to appear in the grazing meadows in the 1930s. Reedbeds then developed and it became a famous wild-fowling area. The colliery closed in 1968 and the reserve was purchased the same year.

1.1.2 Flora and fauna

The ditches which cross the reserve have some interesting aquatic fauna, but the meadows are not floristically rich. The reedbeds are uniform stands of *Phragmites australis* with localised stands of greater pond sedge (*Carex riparia*) and reed mace (*Typha latifolia*). The main conservation interest of the site is ornithological and there are several species of national importance. These include Kingfisher (*Alcedo atthis*), Bearded Tit (*Panurus biarmicus*), Savi's Warbler (*Locustella luscinioides*) and Cetti's Warbler (*Cettia cetti*). There are also Gadwall (*Anas strepera*), Garganey (*Anas querquedula*), Reed Warbler (*Acrocephalus scirpaceus*), Teal (*Anas crecca*), Tufted duck (*Aythya fuligula*), Pochard (*Aythya ferina*), Shoveller (*Spatula clypeata*), Snipe (*Gallinago gallinago*) Hobby (*Falco subbuteo*) and migratory birds including Swallow (*Hirundo rustica*), Sand Martin (*Riparia riparia*), Wagtails (*Motacilla*) and Geese. Bitterns (*Botaurus stellaris*) are also present and the main focus of English Nature's work at present is to encourage Bitterns to return to breed.

1.1.3 Designations

1.1.3.1 Site of Special Scientific Interest

Sites of special scientific interest (SSSIs) are declared by English Nature under the Wildlife and Countryside Act 1981 as "areas of land or water...of outstanding value for their wildlife or geology". Stodmarsh was designated in 1951 because the area contains a wide range of wetland habitats and open water, which support a rich flora and fauna. It is considered a typical southern eutrophic flood plain. It is recognised for its rare plants and invertebrates and also for its abundant and rare bird species as well as the fact that the reedbeds are of an unusual size in Britain

1.1.3.2 National Nature Reserve

Stodmarsh is a National Nature Reserve as declared by English Nature under the National Parks and Access to the Countryside Act 1949. NNRs are established to protect sites of national importance for their biological or geological interest. In the case of Stodmarsh, the reserve is also owned and managed by English Nature. It is managed to prioritise the protection of the habitats and species and also as a resource for scientific research. Each reserve has a management plan, updated every five years. For Stodmarsh NNR management objectives include:

- the management of wetland habitats, especially reedbeds to maintain and enhance populations of rare bird species and associated wildlife,
- the management of reedbeds to provide a varied age structure ensuring suitable conditions for Bitterns, Bearded Tits and Hen Harriers,
- to influence, educate and inspire others by the provision of visitor facilities and management demonstrations,
- to manage public access to protect breeding birds,
- to ensure an adequate supply of water to the reserve.

1.1.3.3 Special Protection Area

Special Protection Areas (SPAs) are designated under the Birds Directive 79/409/EC. The aim is to conserve the habitat of certain rare or vulnerable birds and the habitat of migratory birds. Stodmarsh was designated in 1993 because of its provision of wintering and breeding habitats for wetland bird species. The site qualifies due to its significant populations of Bitterns, Hen Harriers, Gadwall and Bearded Tit and also due to the large range of bird species that are present.

1.1.3.4 Special Area of Conservation

Special Areas of Conservation (SACs) are designated under the Habitats Directive for the protection of habitats and (non-bird) species. Stodmarsh has been designated as a special area of conservation due to the presence of the Desmoulin's whorl snail (*Vertigo moulinsiana*) which lives in the ditches surrounding the wet grassland.

1.1.3.5 Ramsar Wetland of International Importance

Ramsar sites are named after the town in Iran that hosted the original Convention on Wetlands on International Importance Especially as a Waterfowl Habitat in 1971. The status of a Ramsar site is equivalent to that of an SSSI, though its international importance is recognised for planning purposes. Stodmarsh was designated because the site supports a number of rare species, including nationally rare plants, a diverse invertebrate fauna and particularly rare breeding bird species including Gadwall, Garganey, Pochard, Savi's Warbler and Bearded Tit. The site also qualifies by supporting more than 1% of the UK breeding wintering populations of a number of species.

1.1.4 Site management

As the site is vulnerable to drying, the water levels have to be carefully controlled. The site manager controls the water levels to provide a flow of water to all habitats. The Lampen Wall divides the water regime at Stodmarsh into two distinct areas and allows the manipulation of the water levels on the reedbeds without affecting neighbouring farmland. Past floods have, however, resulted in the overtopping of the wall. The water supply for the whole reserve area, in addition to precipitation, is the Lampen Stream. The supply of water for the reedbeds and grazing marsh is controlled from a sluice on the Lampen Stream and there are also two pen stop dams. This allows English Nature to control water levels independently of outside interests. The managers aim to keep the water at a mean depth of 10-25 cm, as this has been deemed to be ideal for Bittern breeding. Within the SSSI cattle are used for summer grazing and the drier fields are cut for silage. Cattle grazing is controlled to prevent trampling of ground nesting birds and to ensure a fairly tightly grazed sward for wintering wildfowl and nesting waders.

Stodmarsh was one of the first NNRs to regularise reed cutting for management (Marren and English Nature 1994). They began to experiment with a reed-harvesting machine in the early 1970s. A rotating double scythe attached to a rotating double frame automatically binds each bundle of reeds for the bailer to stack onto the trailer. Small patches of the reedbed are cut each year to preserve the rest of the bed for nesting birds. The network of open ditches is maintained by regular weed-cutting.

1.2 AIMS AND OBJECTIVES

Increasing agricultural and domestic water supply demands, speculation about climate change and new European directives have raised awareness of the need for accurate estimates of hydrological fluxes. This is particularly true in wetland environments where each of the habitats is tightly linked to water level management to maintain suitable conditions for the species found there. Reduced water tables can cause invasion of species into the ecosystem posing a potential threat to the bird and invertebrate life.

Kent is one of the driest parts of the UK and periods of drought in the mid-1990s resulted in an increased water demand by farmers and consequently a decreased supply to the NNR. There is concern that current water sources may be insufficient to meet future needs. Future climate change has been suggested to be likely to reduce summer rainfall (Hulme *et al.* 2002), which will lower water tables and further increase irrigation requirements. It has therefore been necessary to consider abstracting water from the adjacent River Great Stour or using wind pumps on the feed stream, the Lampen Stream, to augment water supply during low flow conditions. In order to do this, an estimation of the amount of water required by the reserve is necessary.

Quantified estimates of water requirements of, and water availability to, conservation areas are rarely available. There has been little work on the hydrological requirements of natural areas compared to the large amount of research into agricultural crops. This may be due in part to the fact that the EU Habitats Directive prioritises the needs of the environment over those of other user groups which means that owners of protected areas have no need to accurately estimate water requirements as their needs will always be met. This is a source of frustration to people who would like to see ecology and hydrology brought closer together.

The aim of this research is to improve the management of Stodmarsh NNR for the benefit of the wildlife habitat and the species it supports. This will be achieved by increasing understanding of the hydrology of the reserve, a key determinant in the quality of the wetland habitat. The primary objective is to calculate the amount of water that is used by Stodmarsh National Nature Reserve in a year of particular rainfall. This will be done in order to determine whether current water resources are likely to be sufficient in the future, and if not, the quantity and frequency with which water must be extracted from alternative water sources. Within this brief, the primary area of uncertainty within the water balance of the reserve is the rate of evapotranspiration from reedbeds. Therefore a further objective is to increase understanding of the evapotranspiration rates of large reedbeds.

1.3 RESEARCH QUESTIONS

There are two main objectives of this research:

- To calculate the amount of water that is used by Stodmarsh National Nature Reserve in a year of particular rainfall.
- To increase understanding of evapotranspiration (ET) from large reedbeds.

Within these objectives the work breaks down into a number of research questions. These come about from the flow diagram that describes the research in Figure 1.03.

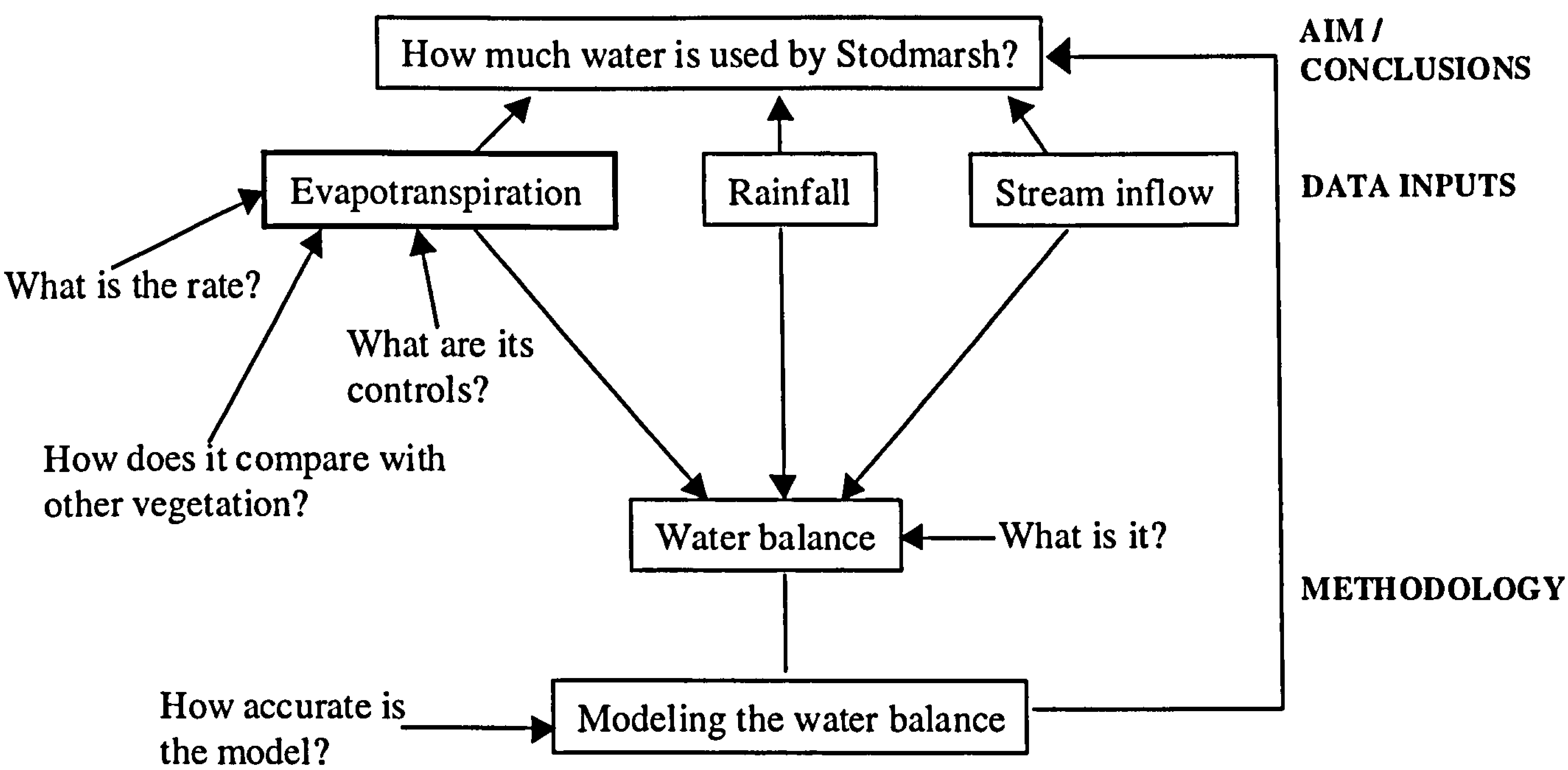


Figure 1.03 – Flow diagram describing the questions that need to be answered in order to meet the objectives of the research.

The structure of the thesis is defined by the overall objective shown at the top of the figure above. The factors that affect the amount of water used by Stodmarsh NNR are considered. From the site description it can be deduced that these will be rainfall amount, the flow of the Lampen Stream and evapotranspiration. These are the data inputs that will be required in order to answer the initial question posed. The former two are comparatively well understood but the latter is poorly understood as related to reedbed environments and therefore an emphasis will be placed on increasing understanding of this part of the water balance. This will require investigating currently uncertain areas such as the magnitude of the evapotranspiration flux from reedbeds, how this compares with other vegetation types using standard evapotranspiration equations and what the controls on evapotranspiration from reeds are. This will be important theoretical research in itself and will also provide an important input into the next stage of the research. All the data inputs will be brought together in order to feed into a water balance. The nature of the water balance at the site will be established and then modelled in order to answer the original aim of the work. As an extension to this, the model input data will be perturbed in order to simulate the potential impacts of climate change on the reserve.

The research questions are listed below and then explained in more detail:

Reedbed evapotranspiration:

1. What is the actual rate of evapotranspiration from reedbeds? (Chapter 3)
2. How can reedbed evapotranspiration be modelled? (Chapters 3 and 4)
3. How does reedbed evapotranspiration compare with that of other surfaces? (Chapters 3 and 4)
4. What are the controls on reedbed evapotranspiration? (Chapter 4)

Water balance creation:

5. What was the site water balance during the study period? (Chapter 5)
6. How can the whole site water balance be modelled? (Chapter 6)

7. What are the water requirements of the site under different weather conditions? (Chapter 6)
8. What is the potential impact of climate change on the site? (Chapter7)
9. How accurate is the water balance? (Chapter 8)

1.3.1 Details of Research Questions

Reedbed evapotranspiration understanding:

1. What is the actual rate of evapotranspiration from reedbeds?

Justification: There is a lack of knowledge of the rate of evapotranspiration from UK reedbeds and disagreement between data sets that have been collected. Much of the work has been done on small reedbeds where advection is important.

Approach: The Bowen ratio energy balance approach was used at Stodmarsh NNR for the summers of 2001 and 2002 to measure evapotranspiration.

Data: 20 minute evapotranspiration data for the growing season of reeds

Equipment: Bowen ratio meteorological instrumentation.

Analysis: Daily average values of evapotranspiration from reeds were found.

2. How can reedbed evapotranspiration be modelled?

Justification: Due to the expense and time consuming nature of taking evapotranspiration measurements, and the lack of long term data, it is necessary to have a methodology for predicting evapotranspiration.

Approach: Two approaches are compared – using crop coefficients as described by Allen *et al.* (1994c) to adjust reference evapotranspiration for reeds and calibrating the Penman Monteith equation directly for reeds.

Data: Crop coefficients are created using the Bowen ratio data described above. The Penman Monteith equation is used with meteorological data, initially from the Bowen ratio equipment and then from a local meteorological station. The Penman Monteith equation is parameterised using data from a wind profile and from stomatal conductance measurements taken using an infra red gas analyser.

Analysis: Models are assessed using scatter diagrams and r^2 values and assessed according to how closely they predict evapotranspiration to that estimated using Bowen ratio approach.

3. How does reedbed evapotranspiration compare with other crops?

Justification: There has been much discussion but few conclusions about the hydrological role of tall wetland macrophytes and how their evapotranspiration compares with terrestrial plants and with open water evaporation.

Approach: Crop coefficients will be used as a means of comparison. Comparisons will also be made using physiological features such as stomatal conductance, atmospheric coupling and aerodynamic resistance.

Data: Bowen ratio data as above for crop coefficients, wind profile data, stomatal conductance measurements.

Analysis: Comparisons of evapotranspiration will be mainly made with a short grass crop similar to the reference surface described by Allen *et al.* (1994c).

4. What are the controls on reedbed evapotranspiration?

Justification: There is a lack of understanding of the controls on reedbed evapotranspiration. Wetland macrophytes have often been thought to act as passive wicks to transpiration.

Approach: Stomatal conductance was measured using an IRGA and estimated using the inverted Penman Monteith equation. The importance of stomatal resistance was compared with aerodynamic factors.

Data: As above

Analysis: Aerodynamic and surface resistance was measured for reeds. The values calculated were compared with those for the reference crop and the omega factor of Jarvis (1976) was used to assess the controls on evapotranspiration.

Water balance creation:

5. What was the site water balance during the study period?

Justification: A water balance is the first step in understanding the hydrology of a wetland site.

Approach: Each component of the water balance is measured and brought together in a water balance equation.

Data: Rainfall data, stream discharge data, gaugeboard readings of water stage, ponded water depth measurements, Bowen ratio evapotranspiration data, Penman Monteith reference evapotranspiration data calculated from a nearby weather station

Analysis: Storage of water on the site was estimated from the water balance model and this was verified using measured storage data.

6. How can the whole site water balance be effectively modelled?

Justification: In order to predict future storage levels on the site under different meteorological conditions to those experienced in the study years, a model is required

Approach: The model was based on the measured water balance and involved using an appropriate ET model, a rainfall runoff model and historical rainfall and meteorological data.

Data: Rainfall and meteorological data from 1973-2002.

Analysis: The model was validated using measured data and run over the historical years to monitor how storage on the site would change in wet and dry years.

7. What is the predicted storage volume in a year of particular rainfall?

Justification: This is the initial question asked at the start of the research

Approach: The water balance model will be run through years of different rainfall amounts to investigate how the site water levels respond.

Analysis: The annual water deficits will be analysed to determine whether additional water would be required to maintain the site in optimal condition.

8. How accurate is the water balance?

Justification: All models have some degree of uncertainty and it is important that this is quantified in order that the degree of confidence with which it can be applied in the future is known

Approach: Errors are calculated both for the measured and modelled water balances by the addition of instrumental errors, Monte Carlo analysis and estimating the accuracy of models.

Data: The measured water balance, water balance model output and manufacturers' instrumental error estimates.

Analysis: The maximum and most likely propagation of instrumental and modelling errors through the water balance equation is calculated in order to put confidence limits on the results.

9. What is the potential impact of climate change on the hydrology of Stodmarsh?

Justification: There is already evidence that climate change is occurring. This is set to be most severe in the South-East of the UK and will impact the water balance, causing potential damage to wetland ecosystems.

Approach: Historical data sets used as water balance inputs will be perturbed according to climate change scenarios.

Data: UKCIP02 climate change scenarios, historical meteorological data

Analysis: Water requirements as predicted by the water balance model with input data perturbed according to climate change scenarios will be compared with data from 1974-2000 in order to predict the potential impact of climate change on the reserve.

1.4 THESIS STRUCTURE

The thesis' structure follows the research questions above. Chapter 2 provides an introduction to the importance of wetland habitats, in particular the issue of wetland hydrology and sets the research in its wider context. Each subsequent chapter then begins with a review of the main literature specifically relevant to that chapter. Chapters 3 and 4 concentrate on the experiments concerned with evapotranspiration from reedbeds. Chapter 3 describes the use of the Bowen ratio kit in 2001 and the results from this (this chapter has been accepted for publication as: Peacock, C.E., Hess, T.M. (In press) Estimating evapotranspiration from a reedbed using the Bowen ratio energy balance method. *Hydrological Processes*). Chapter 4 describes attempts to calibrate the Penman Monteith equation directly in order to create a model of evapotranspiration. The remaining chapters describe how the evapotranspiration models are used with other data to create a water balance model of the site. In Chapter 5 this is done with the measured data collected in 2000 and 2001 and in Chapter 6 the water balance is created using modelled and historical data. Chapter 7 extends the model to include the potential impact of climate change on the site. Chapter 8 discusses the magnitude of errors in the water balance model and Chapter 9 attempts to draw some conclusions.

CHAPTER 2 - Introduction to Wetlands and Wetland Hydrology

2.1 INTRODUCTION

This chapter places the issues at Stodmarsh National Nature Reserve into the wider context of wetland conservation in the UK. It starts by introducing the importance of wetlands and their conservation and then gives some background on the specific habitat types found at Stodmarsh National Nature Reserve – reedbeds and wet grasslands. This study is specifically focussed on hydrology and therefore the importance of wetland hydrology is subsequently introduced followed by more specific information on the hydrology of reedbeds and wet grasslands. The final section of this chapter focuses on the legislation and policy relevant to the conservation of wetland habitats.

2.2 WETLAND HABITATS

Wetlands occupy around 9×10^6 km^2 (around 6%) of the global land area and are an important part of many landscapes worldwide (Acreman 2000). They are often described as representing a continuum between aquatic and terrestrial ecosystems (Brinson 1993a). They include a diverse range of habitat types including floodplains, freshwater marshes, estuaries, coasts, lakes, peatlands and swamp forests. There are more than 50 definitions of a wetland in current use but a typical one is:
"an ecosystem that arises when inundation by water produces soils dominated by anaerobic processes and forces the biota, particularly rooted plants to exhibit adaptations to tolerate flooding" (Keddy 2000 p.3).

Wetlands are some of the most important ecosystems on earth (Mitsch and Gosselink 2000). They have always influenced humans and ancient civilisations grew up along the banks of rivers. However they have also been traditionally thought of negatively, as wastelands of little economic value and harbours of disease such as malaria. Prior to the 1970s, drainage and destruction were accepted policy and actively encouraged. (Dugan 1990; Mitsch and Gosselink 2000). Drainage was mainly for agriculture and urban development. The US has lost 54% of its original wetlands and in the UK in the twentieth century large tracts of wetlands such as the mosses of North-West England and the fens of East Anglia have been turned over to intensive farming (Gilman 1994). However more recently, attitudes have changed and there has been increasing appreciation of the importance of wetland functions. They produce fossil fuels, are important for fish and wildlife protection, they stabilise water supplies, reduce flooding and drought, help erosion control, improve water quality and recharge groundwater aquifers (Mitsch and Gosselink 2000). They are also used as flood storage and in storm flow modification, in groundwater recharge, and can be indicators of available groundwater supplies and are important in erosion control. They also improve the quality of water as it flows through the wetland as they are settling areas for sediment – in some wetlands 80-90% of sediment can settle out and nutrients and trace metals may be removed and retained by the wetland (Carter *et al.* 1978).

2.2.1 Reedbeds

Reedbeds are wetland habitats comprised mainly of the macrophyte *Phragmites australis*. As with many wetland types they are a seral stage in the progression from an aquatic habitat to a terrestrial one. Reedbeds are not species rich environments in terms of vegetation but have high ecological value due to the high biological diversity of the bird and invertebrate life that they support. They provide an important habitat for a number of Red Listed bird species (species in rapid decline as defined by Birdlife International and RSPB) and are also home to over 700 species of invertebrate (Andrews and Ward 1991). There are three nationally scarce birds – Bittern, Bearded Tit and Savi's Warbler, that are entirely dependent on reedbeds, as well as Reed Warbler and Sedge Warbler for whom it is their major habitat. Marsh Harrier and Cetti's Warbler are substantially dependant on reeds (Andrews and Ward 1991). Cetti's Warbler was not found in Britain until 1971 and Stodmarsh was its first known breeding place in 1973. Kent is still its stronghold (Everett 1989). Reedbeds also have economic value and have been observed to be able to remove nitrates from water. Productivity is high and reeds are still used for fencing and thatching today in Romania, Iraq, Japan and China (Mitsch and Gosselink 2000).

The area covered by reedbeds has been declining in this country for the last two hundred years and since 1945, 40% of reedbeds in the UK have been lost (UK Steering Group 1995). They continue to be threatened by the increasing demand for water for domestic and agricultural use. *Phragmites* still occurs throughout the UK but few stands are of sufficient size or have suitable hydrology to have conservation value (Andrews and Ward 1991). Approximately 900 reedbeds make up the surviving 5000 ha in Britain (Brookhouse 1998) but an RSPB survey in 1979-80 showed that there were only 109 reedbeds greater than 2 ha in area and 33 greater than 20 ha. 80% of rare birds that regularly use reedbeds need sites greater than 20 ha in area (Everett 1989). More recently the importance of this resource has begun to be recognised and the UK Biodiversity action plan aims to create 1200 ha of new reedbed by 2010 (UK Steering Group 1995).

One of the reasons for the increase in interest in reedbed creation and conservation is that the fate of the rare Bittern is closely tied up with that of reedbeds. Bitterns disappeared from the UK in the nineteenth century due to hunting but by the 1920s had reinvaded from the continent and were well established again in Norfolk. However they have declined by 70% since 1970. The reasons for the recent decline are unknown but may be due to loss of reeds and reduced water quality. The highest density in Britain is 1 nest per 20 ha, so large sites such as Stodmarsh are very important. A plentiful supply of small eels is also important and these are more likely to penetrate large reedbeds than small ones (Tyler 1994).

2.2.2 Wet grasslands

Stodmarsh is one of 25 internationally important wet grassland sites in the UK. Lowland wet grasslands are defined as being below 200 m in altitude and subject to periodic freshwater flooding or water logging (O'Brien and Self 1994). Wet grasslands support a rich array of wildlife. Across Britain they are home to 500 species of vascular plant and the ditches which surround each field in order to control flooding and drainage support 130 of the 170 species of higher aquatic plants found in the UK. The ditches are also very important for aquatic invertebrates especially dragonflies, water beetles and snails. Eight species of breeding wader are associated with wet grassland including Redshank, Snipe and Lapwing and they are also important for wildfowl. More than 40 bird species of conservation importance are dependant or partially dependant on wet grasslands for breeding and wintering as are 16 species of rare and scarce vascular plant (Fowler *et al.* 1998). 24 red list species are at least partially dependant on lowland wet grasslands (Self *et al.* 1994).

Wet grassland once covered hundreds of thousands of hectares but is one of the most rapidly diminishing wetland types due to agricultural intensification. Since the 1940s around 40% of the habitat has been lost. Of the 1 217 000 ha low lying flood plain land that was, or has the potential to be lowland wet grassland in England and Wales, only 240 000 ha remains as wet grassland, only 109 000 ha holds any breeding waders and only 20 000 ha is agriculturally unimproved (Self *et al.* 1994). In southern areas, including Kent, Lapwing decreased by 68%, Snipe by 26% and Curlew by 50% between 1982 and 1989 (O'Brien and Self 1994). It is one of the most threatened habitats in the UK (Moffat 1994; Self *et al.* 1994) and a priority for conservation. Although recently there has been a decrease in the rate of loss there is still a decline in quality (Fowler *et al.* 1998) and up to 24% could suffer damage by 2019 (Self *et al.* 1994).

The biodiversity action plan targets are to maintain the habitat extent and quality and to rehabilitate 10 000 ha of grazing marsh that have become too dry or intensively managed by 2000 and to create 500 ha of grazing marsh from arable land (UK Steering Group 1995).

2.3 WETLAND HYDROLOGY

An understanding of the hydrology of wetlands is fundamental to an understanding of the entire habitat. Mitsch and Gosselink (2000 p.108) state that "hydrology is probably the single most important determinant for the establishment and maintenance of specific types of wetlands and wetland processes. When hydrological conditions in wetlands change even slightly, the biota may respond with massive changes in species richness and ecosystem productivity." The hydrological regime is supremely important for maintaining the structure and functions of a wetland (Thompson and Finlayson 2001).

A wetland's water level regime is the seasonal pattern of surface and subsurface water elevation. There is great variation in regime between wetland types but constancy in its pattern in one particular wetland ensures reasonable stability. The hydrological regime of a wetland affects chemical and physical properties including soil and water salinity, soil anaerobosis, nutrient availability, pH and sediment properties including deposition rates and texture. Biotic characteristics of the wetland, including vegetation composition and ecosystem structure and function, are also directly affected by hydrology and controlled by factors such as inundation frequency and water depth. Where inundation is greater and for longer, plants require more adaptations to establish, live and reproduce within a wetland. Fluctuations in soil saturation and the subsequent changes in oxidation and reduction conditions exert a strong influence in nutrient cycling, particularly nitrogen. Long term saturation reduces the oxygen content of the soil, limiting the macrophyte species that can survive. Therefore, in general, wetlands with a long flood duration are less diverse. There are also feedbacks between vegetation and hydrology – the growth of vegetation can alter the hydrological regime by slowing water movement and increasing sediment deposition. The accumulation of decaying plant matter combined with trapped sediment can reduce the depth of water in the wetland over time and lead to change in vegetation type.

All natural wetland functions are the result of, or are closely related to wetland hydrology and therefore an understanding of the presence and movement of water through a wetland is vital to an in-depth understanding of wetland function (Carter *et al.* 1978). Hydrological data are required to monitor changes in the water regime and water sources, as a basis for on-site hydrological management (Gilvear and Bradley 2000) and to predict the precise impact of any water management scheme (Acreman 2000). The effectiveness of many management practices is limited because of a lack of understanding of basic wetland hydrological processes (Winter and llamas 1993). This is due to a lack of data and to the traditional neglect of wetlands. Only rarely is hydrological data gathered from wetlands, although their ecology is often monitored carefully (Hollis and Thompson 1998). Due to the small scale of most wetland hydrological studies, the national network of stations for groundwater and streamflow monitoring are generally too sparse to be useful. The hydrology of a wetland can therefore only be assessed on the basis of local data which must be collected for the purpose (Gilman 1992).

2.3.1 Reedbed hydrology

Despite the fact that reeds can grow in water up to 1 m deep and can tolerate falls of water levels of up to 1 m below the surface, control of water levels is of utmost importance in reedbed management. This is because if the reedbed is allowed to dry out, it can lead to the invasion of other species into the ecosystem and a subsequent reduction in its conservation value as a habitat for birds and invertebrates. Naturally, the accumulation of litter will lead to a gradual reduction of water levels and a succession to drier habitat types. Control of water levels is therefore important in reedbed management in order to prevent succession to woodland and to meet the habitat needs of invertebrates and birdlife.

Water level management is the single most important factor in managing a reedbed for wildlife (Ward 1991). It is not sufficient simply to maintain a constant water level throughout the year, as water level variation contributes to habitat diversity, flushes the system of toxins and decreases the rate of sedimentation (Gilman 1998). It is important to achieve a regime that is in keeping with the wildlife objectives of the site and the amount of water available, the habitat needs of the species of wildlife that currently inhabit the bed and those that are desired to be attracted in the future. There are a number of factors that must be held in balance. Some bird species require shallow water for feeding and nesting, whilst it is important to maintain a substantial depth of water throughout most of the year so that the *Phragmites* bed is not invaded by other species. It is necessary to combine the water requirements of the Bittern, which requires wet feeding areas all year round, and those of Bearded Tits which require dry litter to nest but catch prey at water margins. Summer flooding provides Bittern feeding habitat and inhibits scrub invasion but may prevent the development of the carr-reed interface that Cetti's Warbler requires and remove nesting areas for Bearded Tits.

Bitterns are a very rare species whose exclusive habitat is reedbeds. Many reedbeds are managed in order to encourage their breeding. They require a reed water interface for feeding and wet reedbed 10-30 cm deep for breeding and feeding (Andrews and Ward

1991). The Bittern's leg length confines them to water less than around 25 cm deep. They nest in shallow water typically around 10 cm deep amongst reeds. Cetti's Warblers live on the reedbed fringes where the water table is high but flooding does not occur. Savi's Warblers require stands of pure wet reeds together with areas of herbaceous fen vegetation. The specific water level requirements for the birds for which reedbeds are an important habitat are given in the Table 2.01 (Newbold and Mountford 1997). Figures are for the mean water table – seasonality is not considered.

Table 2.01 – Preferred water depths for breeding and feeding for a range of wetland bird species (after Newbold and Mountford 1997).

Species	Breeding	Feeding
Bittern	10-25/30 cm	10-25/30 cm
Marsh Harrier	10-30 cm	10-30 cm
Bearded Tit	-	10-30 cm
Savi's Warbler	-	10-30 cm
Reed Warbler	10-30 cm	10-30 cm
Sedge Warbler	Terrestrial scrub	10-30 cm
Water Rail	0-30 cm	0-15 cm
Teal	0-30 cm	0-20 cm
Shoveler	Terrestrial	0-30 cm
Mallard	Terrestrial – 30 cm	0-35 cm

A variety of water regimes can be implemented on the reedbed. Water may be maintained at high levels all year round, or dropped after the bird breeding season in June and raised again in autumn, or it may be kept low in autumn and kept dry over winter and be re-flooded in the spring. Water levels will naturally drop over the summer when evapotranspiration is greater than rainfall. Management practices may also have an influence; for instance, levels may be drawn down in autumn to allow reed cutting. Once a regime is decided upon, the ability to replicate it from year to year is desirable as sudden changes in water level can be damaging (Haslam 1970). In some cases a consistent water supply can only be guaranteed by pumping water from nearby water courses which requires an abstraction licence.

Hawke and Jose (1996) suggest the following regime for a site with adequate water supply and good control of water distribution. Water levels should be raised on site as soon as winter cutting has finished (late March to early April) to a maximum surface depth of 30 cm. The surface water should be maintained in the range of 5-30 cm throughout the summer. The water should be drawn down to just below the surface from October onwards to facilitate management work, especially cutting. Some wet habitat should be maintained in the winter to provide feeding habitats for Bitterns, but maximum levels on the reedbed should not exceed 1 m. Levels of 30 cm allow Bitterns and other wildlife to use the reedbed for winter feeding.

The guidelines referring to hydrological considerations for reedbed creation appear to be scanty though it seems an obvious stage in the planning of a new reedbed to assess the volume of water it will require. When the new area at Stodmarsh was being developed it was found that there was little helpful literature available. Emerson (1995) in a thesis on wetland design strategy suggests estimating evapotranspiration from values quoted by Smith and Trafford (1976). Hawke and Jose (1996) suggest using 1.4 as a blanket coefficient as, unlike crops, reedbeds often have an unlimited water supply. Campbell (1993) suggests crop coefficients of 1.3 between June and August and 0.5 for the rest of the year.

2.3.2 Wet grassland hydrology

Establishment of suitable water level regimes is central to the management of wet grasslands. They are subject to periodic flooding and have high water tables. Often they have a complex network of drainage ditches that surround and cross them. Management of flooding in wet grasslands started in the seventeenth century, but as pumps were developed, drainage began on a large scale (Self *et al.* 1994). Agricultural intensification saw many areas of wet grassland converted to arable land.

The availability of water is a crucial factor in wet grassland creation and maintaining water supplies of a suitable quality can be a major challenge. It is necessary to understand both the water balance and the requirements of target habitats. Maintaining high ditch water levels for at least part of the year is important. Gowing *et al.* (1998)

demonstrated that water table levels have a strong influence on the plant communities of wet grasslands in Somerset. Water table levels are also crucial for bird species. Waders favour wet grasslands with ditch water levels close to field levels and shallow pools are important feeding areas. Wildfowl are particularly dependent on open water areas.

2.4 POLICY AND PROTECTION OF WETLANDS

This section introduces some of the policy, legislation and tools that influence the conservation of wetlands in the UK today. Some have been mentioned in Chapter 1 as specific designations that affect Stodmarsh but their statutory basis is given in more detail here. Historically, legislation has encouraged the destruction rather than the protection of wetlands. Under the 1973 Water Act nine local authorities controlled water management. The Regional Land Drainage committees, of which the majority of members had agricultural interests, controlled land drainage within a catchment. At a local level Internal Drainage Boards carried out smaller drainage works. These are composed almost entirely of farmers whose influence was proportional to the amount of land they owned. Drainage Boards were in favour of drainage and conversion (Williams 1990). Because management was so fragmented there was little verification of the desirability of drainage or any enquiry into the environmental and ecological effects. Even when SSSIs became established, before 1981 they were lost at a rate of 12% a year because the Nature Conservancy Council (NCC – the predecessor of English Nature) had no legislation to stop their destruction by farmers and there were many battles between farmers and the NCC over wetlands. The passing of the Wildlife and Countryside Act (1981) and subsequent legislation, accompanied by the international conventions and European directives detailed below have created a situation more favourable for wetland conservation.

2.4.1 Ramsar Convention

The Convention on Wetlands (Ramsar, Iran, 1971), was adopted in 1971. The Convention entered into force in 1975 and the UK as a signatory has an obligation to promote the conservation of wetlands especially by establishing nature reserves. Planning policies must be adapted to further the conservation of wetlands designated and listed under the treaty. Designated wetlands must be treated with due care and encroachment across their borders must be avoided. Following designation, sites may only be removed from the list on the basis of "urgent national interest". In the UK Ramsar obligations have been mainly fulfilled through the designation of SSSIs.

2.4.2 Biodiversity Action Plan

Subsequent to the Rio Earth Summit 1992, the UK signed the Convention on Biodiversity. The UK Biodiversity Action Plan has the overall aim of conserving and enhancing biological diversity within the UK and contributing to the conservation of global biological diversity through all appropriate mechanisms. A Biodiversity Steering Group was established to develop costed targets and action plans for key species and habitats. This includes recommendations and advice for land managers, improved site protection, improved or maintained water quality and sensitive development, planning and control. Plans have been proposed for the 116 most endangered species, including the Bittern and Water Vole, both wetland species found in reedbeds, and a number of wetland habitats have been highlighted as being priorities for conservation, including floodplain grazing marshes, fens, reedbeds, raised bog and blanket bog. Specific issues to be addressed for wetlands include habitat fragmentation, falling water tables due to abstraction, water pollution and a lack of technical advice on the management of wetlands for wildlife (UK Steering Group 1995).

2.4.3 Water Framework Directive

The importance of wetland water requirements has been highlighted by the enforcement of the EC Water Framework Directive (2000/60/EC) (WFD). There is a need to avoid long-term deterioration of freshwater quality and quantity. Water is under pressure from the growth in demand for sufficient quantities of good quality water. The aim of the directive is to get all bodies of surface water to "good" status. Wetlands are a central component of the hydrological cycle, performing environmentally and economically important roles. They are therefore likely to become a key element of water basin management under the Water Framework Directive. They will play an important role in the achievement of "good" status in the catchment, requiring accurate data on wetland hydrological processes for optimal management. There is a change of emphasis in resource planning as the needs of natural habitats are considered before those of the domestic and agricultural sectors. The Final Consultation Paper on the Implementation of the EC Water Framework Directive (DETR 2001) states that the directive will establish a framework for the protection of the direct water needs of wetlands. In order to do this, it is necessary to establish what those needs are, which in turn requires knowledge of wetland hydrology.

Wetlands are mentioned only briefly in the text of the WFD but its overall aims will only be possible if the role of wetlands is integrated into the river basin management plan as wetlands are an integral part of the water cycle. Sustainable management of wetlands is a key aspect of river basin management under WFD. The directive will promote sustainable water use based on a long-term protection of water to conserve aquatic ecosystems and the wetlands depending upon them (DETR 2001).

2.4.4 Habitats Directive (92/43/EEC)

The Directive of 1992 on the Conservation of Natural Habitats and of Wild Flora and Fauna (92/43/EEC) (Habitats Directive) principally aims to promote the maintenance of biodiversity. It provides for the creation of a network of protected areas across the EU which are known as Natura 2000 areas. These consist of Special Areas of Conservation (SACs) designated under the Habitats Directive and Special Protection Areas (SPAs) designated under the Birds Directive (1979) which the Habitats Directive builds on. All member states must compile a list of areas containing the habitat types and species listed in the Directive to be designated as SACs. Alongside this, SPAs continue to be designated. There are 168 habitat types and those most at risk are given priority status. The directive also requires all member states to set up an effective system to prevent the capture, killing, injuring or damaging disturbance of endangered species. This legislation has significance for the protection of wetland habitats as there are around nine freshwater wetland habitats specified, as well as a number of specified species found in wetland areas. This is likely to lead to greater protection for many wetland types, in addition to those already protected as SPAs due to their importance for birdlife. SACs and SPAs are protected by an obligation to take steps to avoid the deterioration of listed sites or significant disturbance to them and to assess any plan that is likely to have an impact on the site. The plan is only permitted if it will not affect the integrity of the site. However derogations allow development if there are no alternative solutions or for "imperative reasons of over-riding public interest". To implement these obligations the UK government has created the Conservation (Natural Habitats etc.) Regulations (1994). These regulations are founded on the basis that SACs and SPAs should be protected by provisions similar to those of SSSIs, although planning is subject to stricter regulation.

2.4.5 Wildlife and Countryside Act (1981)

The Wildlife and Countryside Act is the core legislation for nature conservation in the UK. It covers a wide range of issues related to conservation, most of which impact wetlands indirectly. It provides protection from killing or harm of all wild birds and mammals, including wetland species and also provides protection for wild plants. It includes legislation to protects SPAs and for the declaration and protection of NNRs and SSSIs.

2.4.6 Countryside and Rights of Way Act (2000)

The main significance of the Countryside and Rights of Way Act for wetlands is that it imposes a duty on the Secretary of State to notify English Nature of wetlands designated under the Ramsar Convention. English Nature must then notify the relevant local planning authorities, owners and occupiers of the land, the Environment Agency and the relevant water authorities and drainage boards. Protection is given to sites through planning controls and statutory instrument 1994/27/6. The Countryside and Rights of Way Act also gives new prominence to the UK biodiversity action plans and makes it a duty of ministers and local authorities to take into account the 1992 Convention's biodiversity requirements when carrying out their functions. The Secretary of State is required to publish a list of habitat types and species considered to be of principle importance for conservation of diversity (Hughes *et al.* 2002).

2.4.7 English Nature

English Nature is a government agency set up by the Environment Protection Act 1990 and is funded by the Department of Agriculture, Food and Rural Affairs (DEFRA). It has amongst its principal functions the maintenance and management of NNRs, the designation of SSSIs and the giving of advice to the Secretary of State, local authorities and private individuals. It also commissions and supports research that is relevant to nature conservation. It is the principal body responsible for ensuring UK compliance with international environmental treaty obligations, for example, seeking out government designation and protection of Ramsar sites, SPAs and SACs (Garner and Jones 1997).

2.4.8 SSSIs

The legislation regarding SSSIs is now principally in the Wildlife and Countryside Act (1981) although they were originally set up under the National Parks and Access to the Countryside Act 1949. Originally, if the NCC was of the opinion that any land was of special interest due to its flora, fauna, geology or physical features, it was obliged to notify the local planning authorities. The authorities then had a duty, when considering development on or near the site to consult the NCC. They were not obliged to follow the advice of the NCC but it was an important consideration. However this was inadequate to protect the environment, as in particular in 1949 farmers and foresters were not considered a threat to the environment. The Wildlife and Countryside Act improved protection and now English Nature has to notify not only the local authorities but also the Secretary of State and the owners and occupiers of the land. The emphasis however is still on voluntary restraint by landowners rather than clear statutory prohibition and damage to SSSIs still occurs.

2.4.9 Stewardship schemes

European Union funding is available to assist with the protection of areas where wildlife and landscape are of special importance and potentially vulnerable to changes in agricultural practice. These areas are termed Environmentally Sensitive Areas (ESAs). They are chosen to conserve and enhance the natural beauty of the area. In the UK they are governed by the Agriculture Act 1986 which provides DEFRA with the power to designate ESAs. DEFRA then make voluntary ten year agreements with landowners where they are given incentive payments in order to manage the land within the provisions of the agreement which will mean protecting the environment. Several schemes have been set up in areas where wetlands are vulnerable to drainage by agriculture including the Norfolk Broads and the Somerset Levels, both wet grassland areas.

CHAPTER 3 - Estimating Evapotranspiration Using the Bowen Ratio Energy Balance Method

3.1 INTRODUCTION

3.1.1 Publishing information

The material in this chapter has been published as:

Peacock, C.E., Hess, T.M. (in press) Estimating evapotranspiration from a reedbed using the Bowen ratio energy balance method. *Hydrological Processes.*

Some adjustments have been made to the paper's introduction. Material included in the paper which is duplicated in Chapters 1 and 2 of the thesis has been removed and has been replaced by a fuller literature review of previous research on wetland and reedbed evapotranspiration such is necessary in a thesis but is too verbose for a journal paper. This chapter only shows and analyses the data measured in 2001 in order to preserve the integrity of the paper. The data measured in 2002 is given in Appendix A.

3.1.2 Wetland evapotranspiration

Water balance studies (Gilvear *et al.* 1993; Campbell and Williamson 1997; Souch *et al.* 1998; Burba *et al.* 1999) show that evaporation is the largest hydrological flux through many wetland types during the summer months. The ability to estimate accurately the magnitude of this flux will therefore go a long way towards being able to compute the water balance and plan the water resources and regime required for a wetland site. It is, however, one of the most difficult fluxes to quantify.

Despite the large volume of research on wetland evapotranspiration rates there is little agreement over results and generally the process is poorly understood (Crundwell 1986; Souch *et al.* 1996). Campbell and Williamson (1997) and Lafleur (1990b) state that studies of wetland evapotranspiration, over a range of scientific disciplines, have produced a number of conflicting and inconclusive findings due to the use of inappropriate methods and the lack of recognition of the role of transpiring vegetation.

Uncertainties in the measurement of evapotranspiration arise partly from the large number of variables involved. Little is known about the partitioning of energy between canopy transpiration and evaporation from the water surface (Burba *et al.* 1999). There is an on-going debate over whether macrophytes such as reeds transpire at greater than or less than potential rate with results of studies showing both the latter and the former (Lafleur 1990a). Thompson *et al.* (1999) state that recent studies of a range of wetland vegetation types have shown that plant physiological and canopy structural controls can constrain evaporation rates to well below open water potential rates. In a review Ingram (1983) concludes that the rate of evapotranspiration depends on the vegetation type. Eisenlohr (1966) found that hydrophytes reduce evaporation from open water by sheltering the water from wind and inducing shading and concluded hydrophyte transpiration rates can approach those of open water evaporation. Boyd (1987) went further, suggesting that a number of workers have shown that aquatic plants enhance water loss from lake surfaces up to 1.4-2.0 times and state that their own lysimeter study gave enhancement rates of 1.58 – 1.77 times. Conversely, in a previous review of evapotranspiration from swamps, Linacre (1976) said that the presence and nature of vegetation has a relatively minor effect on evapotranspiration rates, compared with the effect of regional climate and the advection of sensible heat. These ratios are very important to resource managers assessing the possible impact of aquatic plant communities on water budgets. The reasons for the variation in results may include poor experimental design and the drawing of conclusions from a limited amount of work. Idso and Anderson (1988) concluded that most papers that claim an increase in evapotranspiration over that of open water come from small tank experiments affected by the oasis effect. Decreases tended to come from experiments conducted within large natural stands. Idso and Anderson (1988) compared the study of Snyder and Boyd (1987) with their own work. They concluded that expansive canopies of short aquatic macrophytes increase evapotranspiration but short vegetation decreases it.

Theories as to why hydrophyte evapotranspiration may exceed that of open water include the fact that they increase the evaporative surface area due to the foliage and that plants may reduce aerodynamic resistance. It has been suggested that hydrophytes have little control over their stomata and as they live in wet areas must be able to transpire at potential rates. However Anderson and Idso (1987) found evidence that hydrophytes do exhibit stomatal regulation even under well watered non-stressed conditions. They concluded that whether water loss rates are greater or less than those from open water surfaces is a function of the overall geometry of the plant stand.

3.1.3 Reedbed evapotranspiration studies

Like many large wetland plants, *Phragmites australis* is often accorded high rates of transpiration with little experimental verification (Gilman and Newson 1983). Haslam (1970) quotes values of 1000-1500 mm per annum for *Phragmites* in East Anglia. There have, however, been a limited number of studies carried out on evapotranspiration from UK reedbeds. Gilman and Newson (1983) investigated *Phragmites* in an Anglesey wetland and found mean evapotranspiration rates of 2.6 mm d^{-1}. This figure is low compared to those measured in other studies. Fermor *et al.* (2001) have carried out the largest published reedbed study in the UK using small lysimeters installed within *Phragmites* stands. Peak evapotranspiration was recorded between June and August with a maximum of 5.11 mm d^{-1} at Teeside and 4.73 mm d^{-1} in the West Midlands. Crop coefficients were calculated by comparing lysimeter measured evapotranspiration with pan measured evapotranspiration and evapotranspiration as calculated by MORECS (Meteorological Office Rainfall and Evaporation Calculation System). They found that in April *Kc*s were slightly less than 1 due to the sheltering effect of dead reed stems early in the growing season and rise to a peak in late summer or early autumn.

Fermor *et al.* (1999) carried out a review of studies on reedbed evapotranspiration world-wide. It was found that the majority were carried out in central Europe and most used lysimeters or cut stem weight change methods, both of which suffer greatly from problems of advection and change in transpiration rates due to disturbance. Examples of these include Gel'bukh (1964) who measured evapotranspiration from reedbeds in a river delta in Northern Kazakhstan using transpiration evaporimeters. Evaporation ranged from 0.2-11.3 mm d^{-1} and generally reed evaporation was greater than open water evaporation. Krolikowska (1971) measured reedbeds in a lake in Poland using cut shoots which were immediately weighed and then re-weighed 5 minutes later to calculate the amount of water transpired. Herbst and Kappen (1999) researched reedbelts along a narrow lakeshore in Northern Germany. They used the Shuttleworth and Wallace (1985) model which is based on the Penman Monteith equation but considers soil and vegetation evaporation separately. They found transpiration rates in the stand of around 10 mm –15 mm d^{-1} and latent heat was more than twice as high as radiation input indicating significant advection. Reed evapotranspiration always exceeded lake evapotranspiration. These are much higher values than reported in other studies but this is likely to be due to the small area of the reeds studied.

There have been studies carried out using Bowen ratio energy balance technique (BREB) on reedbeds. Burba *et al.* (1999) studied energy fluxes of *Phragmites australis* in prairie wetlands in Nebraska, USA. A marsh of 125 ha with reeds growing in standing water of a depth of 0.5m was studied. It was found that evapotranspiration was the major use of incoming energy and during early and peak growth used 80-90% of net radiation. Evapotranspiration was between 2.5 mm and 6.5 mm per day with an average of 3.75 mm between June and October. There were sharp decreases in evapotranspiration on cloudy or rainy days. They found that during early and peak growth stages actual evapotranspiration was 75-100% of potential but during senescence it was only 10-75%.

Smid (1975) used the Bowen ratio approach to look at vigorous stands of *Phragmites* in Czechoslovakia. During the peak season, sensible heat is small but becomes much more significant during senescence. Daily measured totals of evapotranspiration were 5.6 mm, 6.9 mm, 5.5 mm and 1.4 mm. The results are significantly higher than those found by Fermor *et al.* (1999). This may be due to the fact that evapotranspiration was only measured on the days when the weather favoured high evapotranspiration or due to the fact that Czechoslovakia has a continental climate whereas the UK climate is maritime.

Burgoon *et al.* (1997) estimated evapotranspiration from a created wastewater treatment reedbed in Washington where temperatures are similar to the UK but rainfall is much lower. It was estimated by metering all the water applied to the reed beds and all filtrate water pumped out. Average ET from May to June 1996 was 6.4 mm d^{-1}.

3.1.4 Evapotranspiration modelling

The difficulty in measuring evapotranspiration directly has led to the development of physically based models that allow its estimation from readily available weather data. These calculate ET_c, which is "evapotranspiration from disease-free, well-fertilized crops, grown in large fields, under optimum soil water conditions, and achieving full production under the given climatic conditions" (Allen *et al.* 1998 p. 7). ET_c varies according to meteorological variables – primarily radiation, vapour pressure deficit and windspeed – and also according to canopy characteristics – leaf surface area, the plant height and roughness, reflection and ground cover of the leaves. This second group of factors change with season and growth stage and are therefore difficult to quantify. For this reason, evapotranspiration equations are rarely used in a single step. Instead a crop coefficient approach is used (Pereira *et al.* 1999). The crop coefficient (Kc) is the ratio of ET_c to the evapotranspiration of a reference surface (ET_{ref}) and represents an integration of the effects of the canopy characteristics that distinguish the crop from the reference surface (Allen *et al.* 1998). In this study, ET_{ref} was defined as the evapotranspiration of a surface similar to short green grass with an assumed height of 0.12 m, a fixed surface resistance of 70 s m^{-1} and an albedo of 0.23, following Allen *et al.* (1994c). ET_{ref} is, therefore, a function of the weather only.

Using a crop coefficient removes the need for a separate evapotranspiration equation or measurement for each crop type and season. Once coefficients have been developed they can be used to estimate ET_c using Equation 3.01.

$$ET_c = Kc\, ET_{ref} \quad (3.01)$$

It is assumed that measured evapotranspiration from reeds (ET_{reed}) is equal to ET_c as defined above, as the soil was constantly saturated and the reeds in good condition, and therefore can be used to estimate crop coefficients.

This crop coefficient technique can be very useful in terms of management of wetlands, as direct measurements of evapotranspiration are labour intensive and costly and therefore can only be used over short periods. Ultimately the data collected need to be utilised to create models that can be used over larger spatial and temporal scales. Currently in planning new reedbeds in the UK or in managing the hydrology of existing reedbeds a crop coefficient of 1.4 and the use of Penman potential evapotranspiration estimates from Smith and Trafford (1976) is suggested (Hawke and Jose 1996; Bardsley *et al.* 2001). However there is little basis for this due to the paucity of information on reedbed evapotranspiration.

3.1.5 The Bowen ratio method

In this study ET_{reed} was estimated using the Bowen ratio energy balance method. This is an extensively used method for quantifying evapotranspiration (Heilman *et al.* 1989). Although it is not without problems, and some would argue that it has been superseded by the eddy correlation technique in recent years, it remains a cheaper alternative, which can be more robust, particularly in wet conditions. Thompson *et al.* (1999) and Tanner and Greene (1987) both compared the eddy correlation and Bowen ratio methods and found that there was generally good agreement between the systems.

The Bowen ratio technique has advantages over the commonly employed lysimeter technique, perhaps the most common method of measuring actual evapotranspiration. BREB can provide good estimates of evapotranspiration over larger areas than is possible with lysimeters and avoids the problems of the "oasis" and "clothesline" effects, as well as issues related to the unrepresentitiveness of the disturbed vegetation in lysimeters of the stand. Pruegar *et al.* (1997) compared lysimeters with the Bowen ratio method and found that both methods showed similar seasonal trends and responses to changes in daily meteorological conditions.

There have been a number of studies carried out on wetlands using the Bowen ratio approach. BREB has been used in several studies in Arctic wetland environments (Roulet and Woo 1986; Lafleur 1990a; Lafleur and Roulet 1992; Mendez *et al.* 1998), in peat bogs in a variety of climates (Kim and Verma 1996; Campbell and Williamson 1997; Phersson and Pettersson 1997; Thompson *et al.* 1999) and in freshwater marshes in the USA (Dolan *et al.* 1984; Souch *et al.* 1998), Africa (Rijks 1969) and central Europe (Priban and Ondok 1985), as well as the studies of Smid (1975) and Burba *et al.* (1999) on reedbeds mentioned above. The only published use of Bowen ratio on a wetland in UK has been the work of Gavin and Agnew (2000) on wet grassland in the North Kent marshes.

3.2 OBJECTIVES

The objective of the research described in this chapter is to measure evapotranspiration from reeds in order to develop crop coefficients from the Penman Monteith reference equation that can be used to estimate actual evapotranspiration for management of the study site. The Bowen Ratio Energy Balance approach was used to measure actual evapotranspiration from reedbeds which, whilst having its own limitations, can produce accurate results when the technique's assumptions are met.

3.3 MATERIALS AND METHODS

3.3.1 Study site

The experimental work was carried out at Stodmarsh NNR, Kent, UK (51° 19'N, 1°12'E, elevation <5 m). The site is 243 ha in area and has a fairly simple hydrology with inputs from streamflow and precipitation. It is underlain by clay and therefore groundwater percolation is considered insignificant. The outflows of water are stream outflow and evapotranspiration. The water levels on the site are controlled by a series of bunds and sluices. The instruments were set up in the reserve between 24th May and 28th August 2001. The reeds were in their sixth growing season and their maximum height ranged from 1.54 m on 06/06/01 to 2.16 m on 15/08/01. The leaf area index, measured with a Sunscan Canopy Analysis System (Delta T Devices Ltd, UK), was 3.39 in June, 5.51 in July and 5.71 in August. The soil was permanently saturated with variable depths of ponded water up to 0.17 m.

3.3.2 Bowen ratio theory

The Bowen ratio energy balance method estimates actual evapotranspiration by calculating the partition of convective fluxes between latent and sensible heat (Oke 1993). It is based on the energy balance equation:

$$R_n - G = H + \lambda E \qquad (3.02)$$

where R_n is net heat gain from radiation (W m^{-2}), G is ground heat loss (W m^{-2}), H is sensible heat loss (W m^{-2}) and λE is latent heat loss (W m^{-2}). The latent heat loss is calculated from a rearrangement of Equation 3.02:

$$\lambda E = \frac{(R_n - G)}{(\beta + 1)} \tag{3.03}$$

incorporating the Bowen ratio (β):

$$\beta = \frac{H}{\lambda E} \tag{3.04}$$

Within a few metres of the surface H and λE may be expressed as:

$$\lambda E = \frac{-\rho a C p}{\gamma} k_V \frac{\Delta e}{\Delta z} \tag{3.05}$$

$$H = -\rho C p k_H \frac{\Delta T}{\Delta z} \tag{3.06}$$

where Δe is the change in vapour pressure with height (kPa), ρ is atmospheric density (kPa), Cp is the specific heat of air (kJ/kg K), ΔT is the change in temperature with height (K), z is height (m), γ is the psychrometric constant (Pa K^{-1}) and k_H and k_V are turbulent transfer coefficients for vapour and heat ($m^2 s^{-1}$).

The Bowen ratio is calculated from gradients of temperature and vapour pressure measured at two heights above the canopy by combining equations 3.05 and 3.06. Generally k_H and k_V are not known but are assumed to be equal and therefore cancel out of the equation:

$$\beta = \gamma \frac{\Delta T}{\Delta e} \tag{3.07}$$

The assumptions made when using the Bowen ratio are (Angus and Watts 1984):

- Turbulent transfer coefficients for heat and water vapour are identical. A recent study (Heikinheimo *et al.* 1999) found that the similarity theory applied in all atmospheric stability situations thus confirming earlier work such as that of Businger *et al.* (1971) and Crawford (1965). McNaughton and Laubach (1998) in their similarity studies conclude that Bowen ratio measurements will be satisfactory except with quite intense inversions and that using the assumption will only result in minor error.
- The two levels at which temperature and humidity are measured must be within the layer of the airflow that has adjusted to that surface so that there is an absence of

horizontal gradients of temperature and humidity. This assumption is assessed by estimating required fetch.

3.3.3 Equipment

The Bowen ratio system (Campbell Scientific, Inc. UK) used was similar to that described by Tanner and Greene (1987) and Bingham *et al.* (1987). A photograph of the equipment set up in the reedbed can be seen in Plate 3.01 at the end of this section.

Vapour pressure and temperature were measured at 2.5 m and 4.5 m above the ground at the end of arms, perpendicular to the steel tripod on which they were supported. Net radiation was measured just above the maximum height of the canopy.

Soil heat flux was measured with two soil heat flux plates (Hukseflux Thermal Sensors, Netherlands). These were placed 0.08 m below the soil surface and thermocouples to measure the change in soil temperature were at 0.02 m and 0.06 m below the surface. Heat flux in the soil and water above the surface was estimated following Allen *et al.* (1994a):

$$G = G_{plate} + \frac{\Delta T_s (C_s z_s + C_{sw} z_{sw})}{\Delta t} \qquad (3.08)$$

where G_{plate} is the soil heat flux measured by the heat flux plates (W m^{-2}), ΔT_s is the change in soil temperature between the thermocouples (°C), C_{sw} is the specific heat capacity of water (4.187 MJ m^{-3} °C^{-1}), z_s is the depth of the soil layer being measured (0.08 m), z_{sw} is the depth of the standing water (m), Δt is the time interval (1200 s) and C_s is the specific heat capacity of soil (MJ m^{-3} °C^{-1}), estimated from De Vries (1963):

$$C_s = (1.93F_m + 2.51F_o + 4.187F_w) \qquad (3.09)$$

where F_m, F_o and F_w are the fractions of mineral, organic matter and water in the soil. Mineral and organic matter fractions were calculated as 0.47 and 0.08 respectively (see Appendix B for the methodology of this calculation). Soil moisture content was estimated from spot measurements taken on a weekly basis with a Theta Probe™ (Delta T Devices, UK). As the soil was saturated this showed no systematic change and an average value of 0.45 was used, thus C_s was estimated as 3.12 MJ m^{-3}°C^{-1}. Water depths were on average 0.1 m and had a maximum depth of 0.17 m. There is some question as to the appropriateness of the use of soil heat flux plates in saturated soil due to the difference in thermal conductivity between the plate and the soil. However the thermal conductivity of the soil was measured (Appendix C) and it was concluded to be within the range recommended by the manufacturer.

Weekly maintenance of the Bowen ratio equipment was carried out which involved cleaning the mirror and adjusting the bias on the dewpoint hygrometer. In order to reduce problems of condensation the system was switched off at night (the pump drawing air into the hygrometer was programmed to switch off when net radiation was less than 0 Wm^{-2}). The box containing the dewpoint hygrometer was heated using a 21 W light bulb programmed to illuminate between 06:00 and 09:00 and the box was insulated with polystyrene to prevent the temperature within the box falling to dewpoint temperature.

A problem specific to using the Bowen ratio approach over reeds, as opposed to over shorter vegetation is the reduced flexibility in the height of the arms. The lower arm must be higher than the vegetation that is being sampled, so that it is sampling the bulk canopy surface and not a smaller microclimate within the canopy. The top arm must be low enough to ensure that it is within the lower part of the boundary layer in equilibrium with the canopy surface (as this is an assumption of BREB) and that the fetch requirements are met. At the same time the arms should be as far apart as possible in order to get the best possible resolution of the temperature and vapour pressure gradient data. The height of the reeds means both arms have to be very high resulting in difficulties with maintenance. In order to increase the difference between the arms the tripod was extended in length compared to the standard tripod.

More detail about the practical use of the Bowen ratio system can be found in Appendix D.

Windspeed was measured using a switching anemometer (Vector Instruments, UK) 2.48 m above the ground. Daily rainfall data for West Stourmouth, Kent (around 1 km from the site) and cloud cover and wind direction data from Manston, Kent (around 10 km from the site) were supplied by the British Atmospheric Data Centre.

3.3.4 Fetch Requirements

The fetch requirements for BREB are often suggested to be 100 times the height of the upper sensor (Angus and Watts 1984; Heilman *et al.* 1989; Stannard 1997). However the fetch requirements are also a function of surface roughness (Brutsaert 1982). If it is assumed that the lower 10% of the internal boundary layer is in equilibrium with the surface, then the minimum fetch requirement for near neutral conditions, x_f, is (ASCE 1996 after Gash and Stewart 1986):

$$x_f = \left(\frac{30(z-d))}{(z_o)^{0.125}} \right)^{1.14} \qquad (3.10)$$

where z is maximum sensor height (4.125 m), d is the zero plane of displacement (m) and z_o is the surface roughness length of momentum (m). d and z_o were calculated using the equations given by Brutsaert (1982):

$$d = 0.67\,h \qquad (3.11)$$

$$z_o = 0.123\,h \qquad (3.12)$$

where h is crop height. Fetch lengths were measured in eight directions from the BREB station.

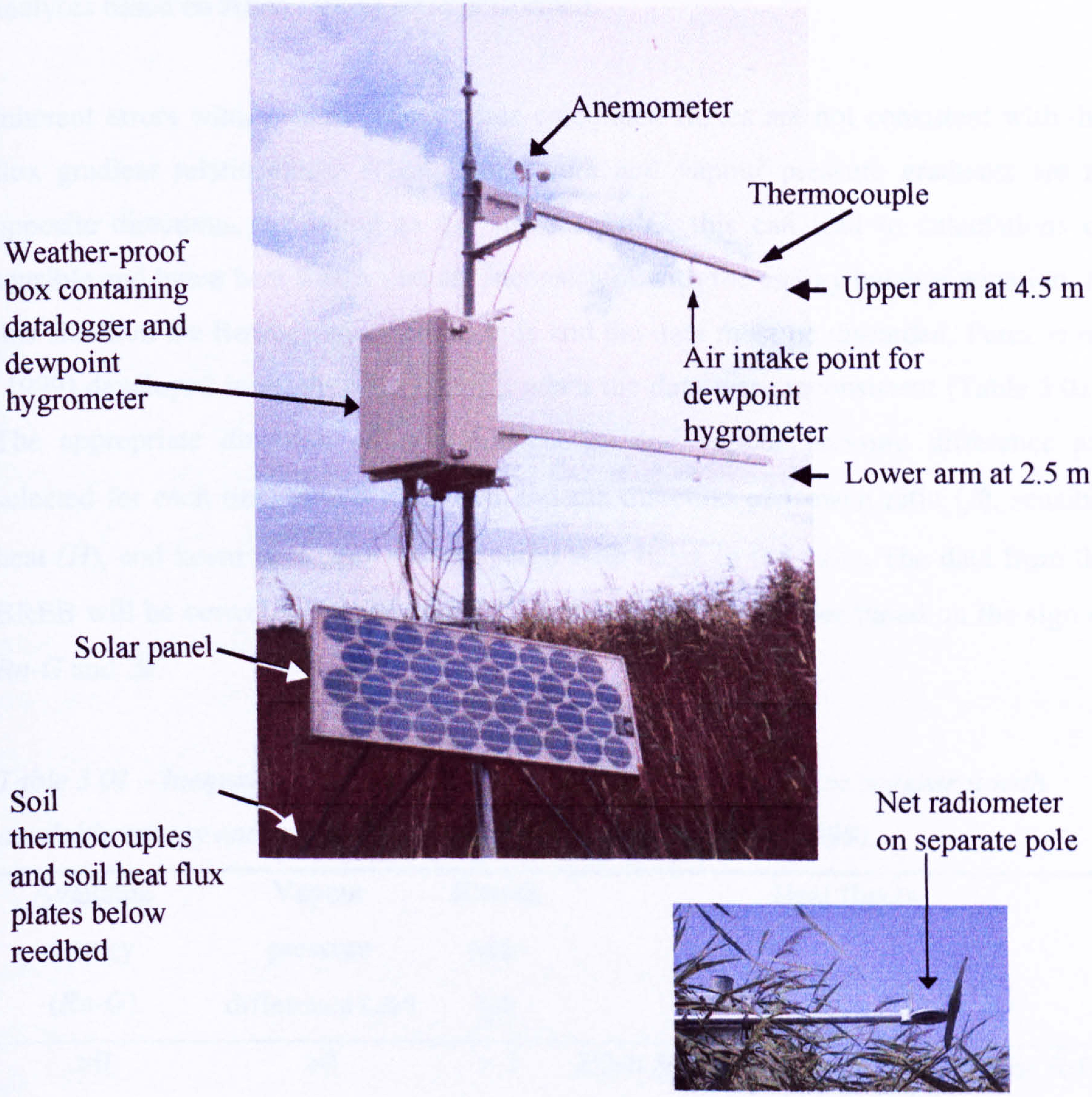


Plate 3.01 – The Bowen ratio energy balance equipment set up on the tripod in the field

3.3.5 Screening of Data

The BREB approach relies on accurate estimates of the Bowen ratio, and this in turn relies on accurate measurement of temperature and vapour pressure gradients. Because of this it was very important to screen the data carefully. Graphical and qualitative analyses based on Allen (1996) were performed.

Inherent errors within BREB occur when calculated fluxes are not consistent with the flux gradient relationships. When temperature and vapour pressure gradients are in opposite directions, according to the sign of *Rn-G*, this can lead to calculations of sensible and latent heat fluxes that are inconsistent with the energy balance equation. In this situation the Bowen ratio method fails and the data must be discarded. Perez *et al.* (1999) developed inequalities to identify when the data were inconsistent (Table 3.01). The appropriate direction of available energy and vapour pressure difference are selected for each time period measured and the direction of Bowen ratio (β), sensible heat (H), and latent heat (λE) are compared with those in the table. The data from the BREB will be correct when they fulfil the appropriate inequalities based on the sign of *Rn-G* and Δe.

Table 3.01 – Inequalities used to determine whether heat fluxes are consistent with available energy and vapour pressure differences (Perez et al. 1999).

Available energy (*Rn-G*)	Vapour pressure difference (Δe)	Bowen ratio (β)	Heat fluxes
>0	>0	>-1	$\lambda E>0$ & $H\leq 0$ for $-1<\beta\leq 0$ or $H>0$ for $\beta>0$
>0	<0	<-1	$\lambda E<0$ and $H>0$
<0	>0	<-1	$\lambda E>0$ and $H<0$
<0	<0	>-1	$\lambda E<0$ and $H\geq 0$ for $-1<\beta\leq 0$ or $H<0$ for $\beta>0$

Errors also occur when β is close to –1. This causes Equation 3.03 to tend towards infinity and occurs when the values of latent and sensible heat are of nearly equal magnitude but with opposite signs. This usually happens near sunrise and sunset due to the small gradients in temperature and vapour pressure and changing flux direction (Pruegar *et al.* 1997). The range of Bowen ratio around –1 that should be excluded depends on the vapour pressure gradient and the resolution limits of the sensors. It can be found with an error analysis of β. The excluded interval can be determined from Perez *et al.* (1999):

$$\varepsilon = \frac{\delta\Delta e - \gamma\delta\Delta t}{\Delta e} \tag{3.13}$$

where $\delta\Delta e$ is the resolution limit of the dewpoint hygrometer (0.02 kPa), $\delta\Delta t$ is the resolution limit of the thermocouples (0.02°C) and ε is the error interval around –1. As a consequence, all data sets with Bowen ratios between 0 and –1.33 were excluded.

3.3.6 Reference evapotranspiration

The Penman Monteith equation was used to create estimates of reference evapotranspiration following Allen *et al.* (1994b):

$$\lambda E = \frac{s(R_n - G) + \rho CpD / r_a}{s + \gamma(1 + r_c / r_a)} \tag{3.14}$$

where s is slope of the saturation vapour pressure curve (kPa° C^{-1}), D is the vapour pressure deficit (kPa), r_c is the canopy resistance (70 s m^{-1}) and r_a is the aerodynamic resistance (s m^{-1}) calculated as:

$$r_a = \frac{\ln\left(\frac{z_m - d}{z_o}\right)\ln\left(\frac{Z_h - d}{z_{oh}}\right)}{k^2 u_z} \tag{3.15}$$

where d is zero plane of displacement (0.08 m), z_o is the roughness parameter for momentum (0.015 m) z_{oh} is the roughness parameter for heat and water vapour (0.0015 m), z_m and z_h are the heights of windspeed and air temperature measurements, k is Von Karman's constant (0.41) and u is windspeed m s^{-1}.

3.3.7 Analysis

Evapotranspiration rates for each 20 minute period in mm day^{-1} were calculated from the 20 minute latent heat data computed from BREB (λE) using Equation 3.16:

$$ET = \left(\frac{86400\lambda E}{1000000}\right)\left(\frac{1}{2.45}\right) \tag{3.16}$$

As vapour pressure was not measured at night, the average daylight value of evapotranspiration from the BREB approach had to be multiplied by the number of daylight hours and divided by 24. Days where more than three hours of data had been removed in screening were missed out entirely and remaining gaps were interpolated using surrounding data.

3.4 RESULTS

Monthly summaries of meteorological parameters for the summer of 2001 are shown in Table 3.02. June had the highest average net radiation (270 W m^{-2}) and the lowest total rainfall (27.6 mm). August had the lowest average net radiation (210 W m^{-2}) and highest total rainfall (86.4 mm).

Table 3.02 –Monthly meteorological data for Stodmarsh NNR June-August 2001

	Av. daily radiation (Wm^{-2})	Av. daily temperature (°C)		Av. daily saturation deficit (kPa)	Total rainfall (mm)	Number of rain days
		max	min			
June	270	18.9	10.0	1.84	27.6	8
July	249	22.5	12.6	0.36	64.6	12
August	210	21.9	13.4	2.19	86.4	15

Figure 3.01 shows the percentage of unusable Bowen ratio data found on a weekly basis.

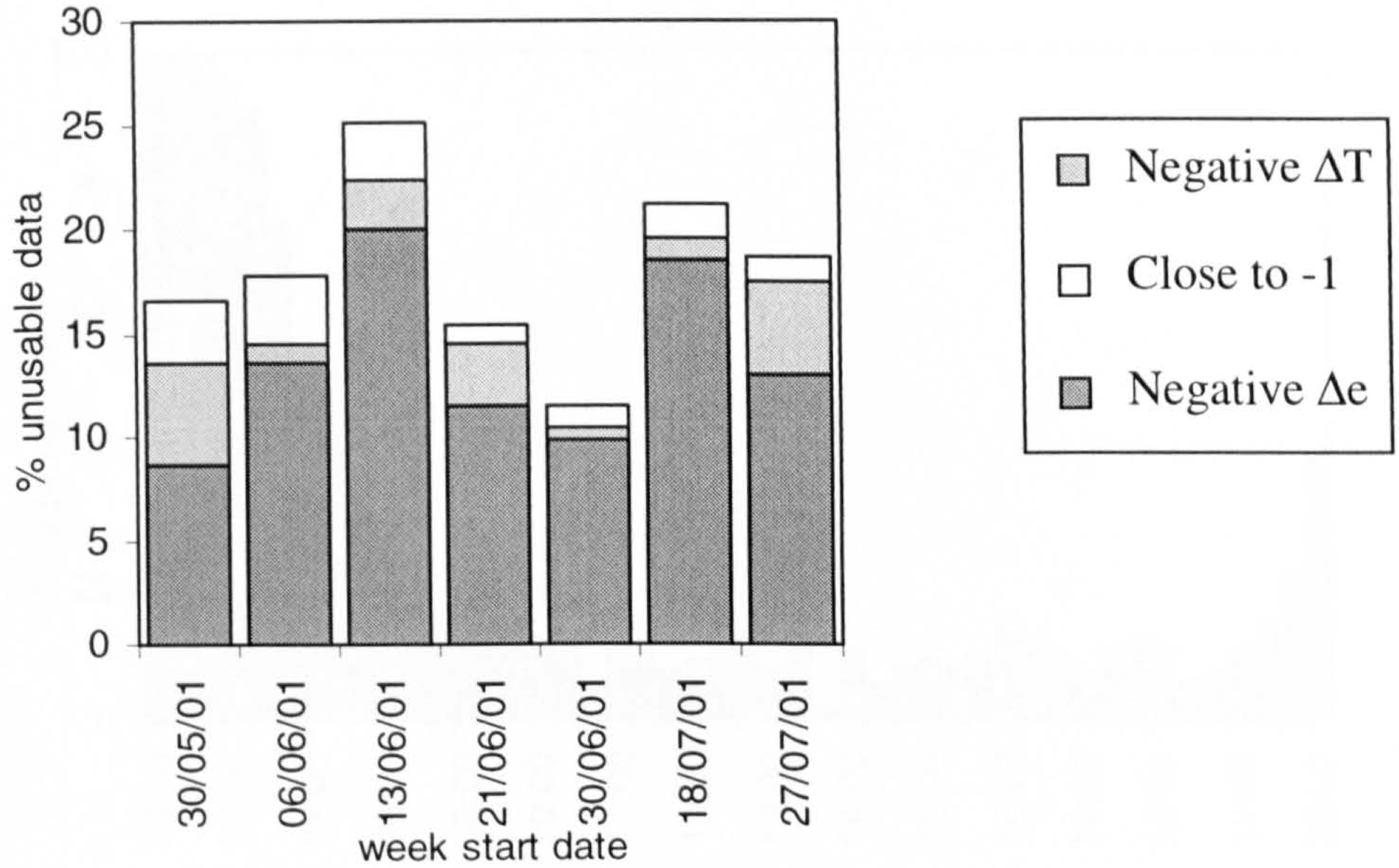


Figure 3.01 – Weekly percentage of unusable data due to flux inconsistencies or the Bowen ratio being close to –1.

The figure was around 20%, which compares well to other authors (e.g. Perez *et al.* 1999). Increased vapour pressure with height in the daytime accounted for 11% of the inconsistencies, and the Bowen ratio being close to –1 accounted for 12%. The greatest cause of inconsistency was an increased temperature with height in the daytime, which accounted for 77% of the unusable data. The majority of errors occurred early in the morning (Figure 3.02) when inconsistencies occur as fluxes change direction, and direction change does not necessarily occur for each flux simultaneously. However the early morning is also a time when evapotranspiration fluxes are small so the loss of data has minimal impact on daytime totals.

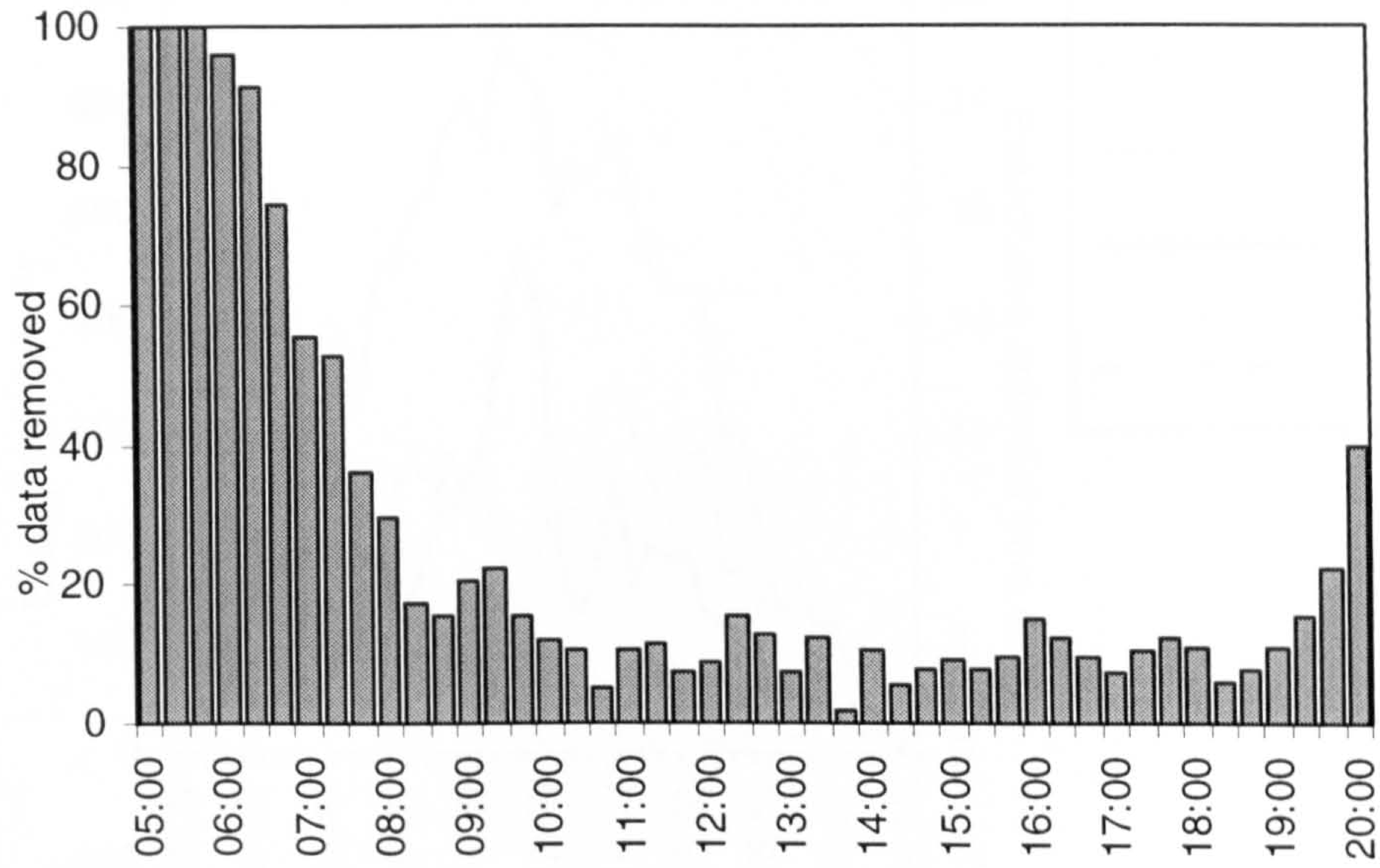


Figure 3.02 –The percentage of the BREB data removed at each 20 minute period of the day throughout the measurement period. The greatest removal occurs in the morning.

For an average maximum crop height of 1.9 m the fetch requirement was 196 m in all directions. The minimum fetch was in a north-westerly direction at 200 m. However, north-westerly winds only comprised 12% of the total. The most common wind direction was south-west, which had a fetch of 833 m. Although the calculated fetch requirement is for near neutral conditions, in cases of instability, which occur quite frequently during the sampled period, the exponent (1.14 – Equation 3.10) is decreased resulting in a reduced fetch requirement. Although stability would result in an increase in fetch requirements, no stable cases occurred in the daytime measurements. (Atmospheric stability was assessed using the Richardson number – see Appendix E).

Figure 3.03 shows energy fluxes over a typical day.

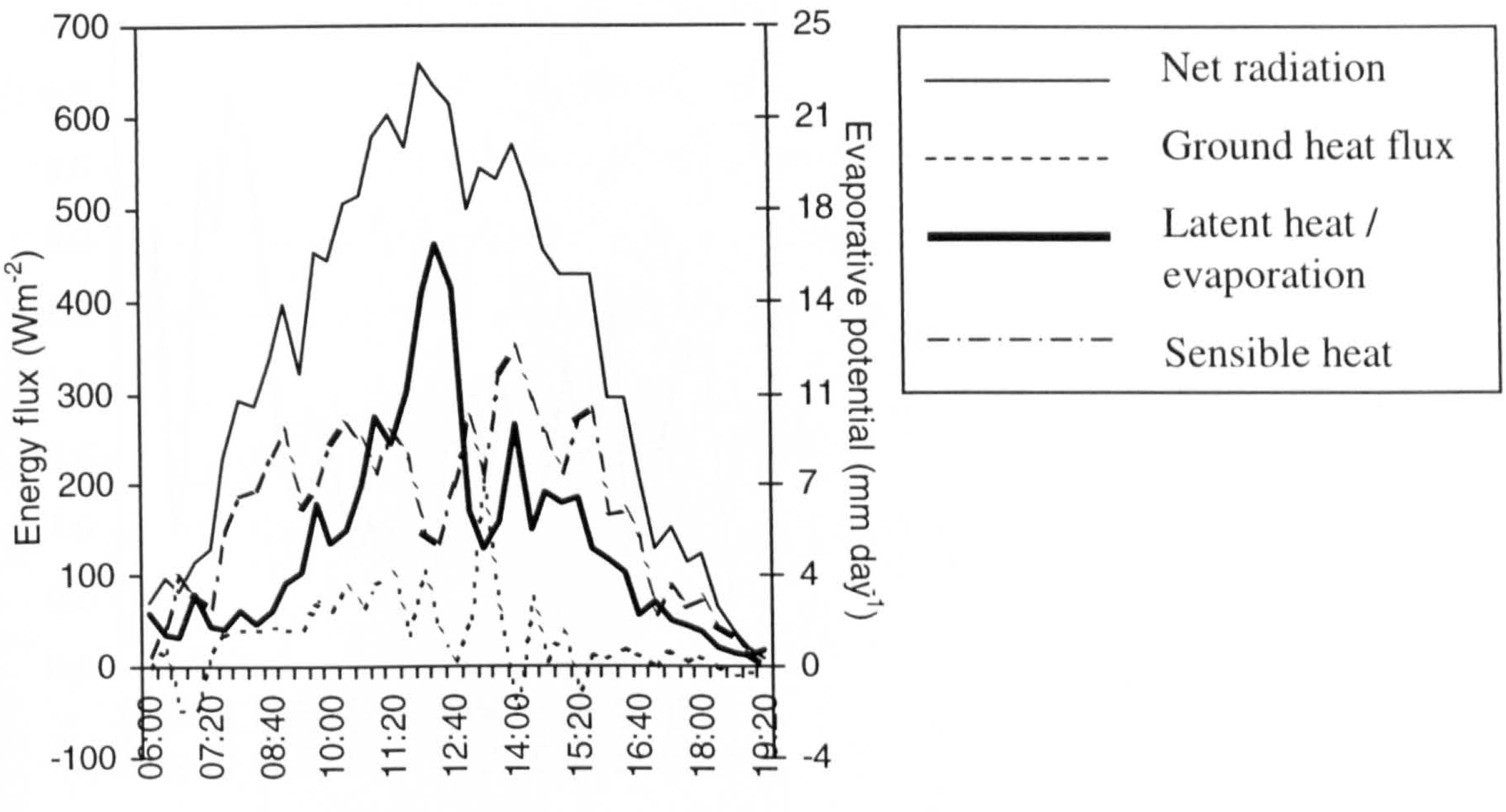


Figure 3.03 – Energy fluxes for 26th June 2001 calculated using the Bowen ratio energy balance approach. The net radiation can be seen together with its division into sensible, latent and ground heat flux. All fluxes may be read on either axis – energy flux magnitude in Wm^{-2} *or evaporative potential of the energy flux in mm* d^{-1}.

Ground heat flux is small compared to net radiation and becomes positive one to two hours after sunrise. Its pattern and magnitude is similar to that measured by Burba *et al.* (1999). In the early morning sensible heat flux is greater than latent heat flux, but latent heat flux becomes larger around midday and peaks at the same time as net radiation. Latent heat flux declines again in the afternoon to become smaller than sensible heat flux. Latent heat flux always remains smaller than net radiation. Evapotranspiration increases during the morning to reach peak rates around midday and then decreases in the afternoon. A plot of the Bowen ratio is also presented for the same day in Figure 3.04. It can be seen that it is highest in the morning and the afternoon and declines around midday.

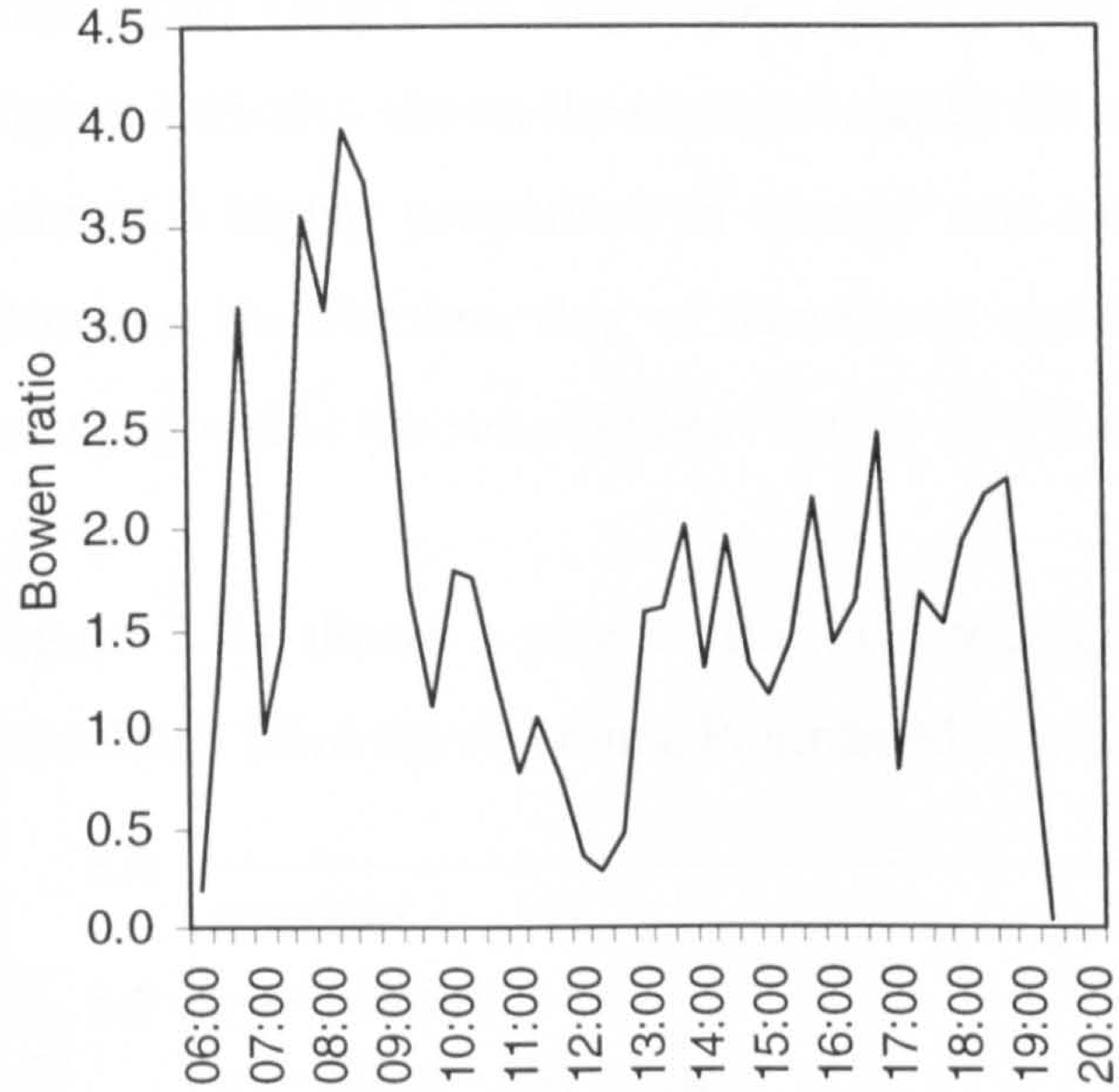


Figure 3.04 – Plot of the Bowen ratio as measured by the Bowen ratio energy balance method for 26/06/02

The average contribution of latent heat flux, ground heat flux and sensible heat flux to the energy balance was calculated for each day (note that these are day light values – the overall contribution of ground heat flux is likely to be less as it is negative at night). The average values are 60% of net radiation used in sensible heat flux, 32% used in latent heat flux and 8% used in ground heat flux (Figure 3.05).

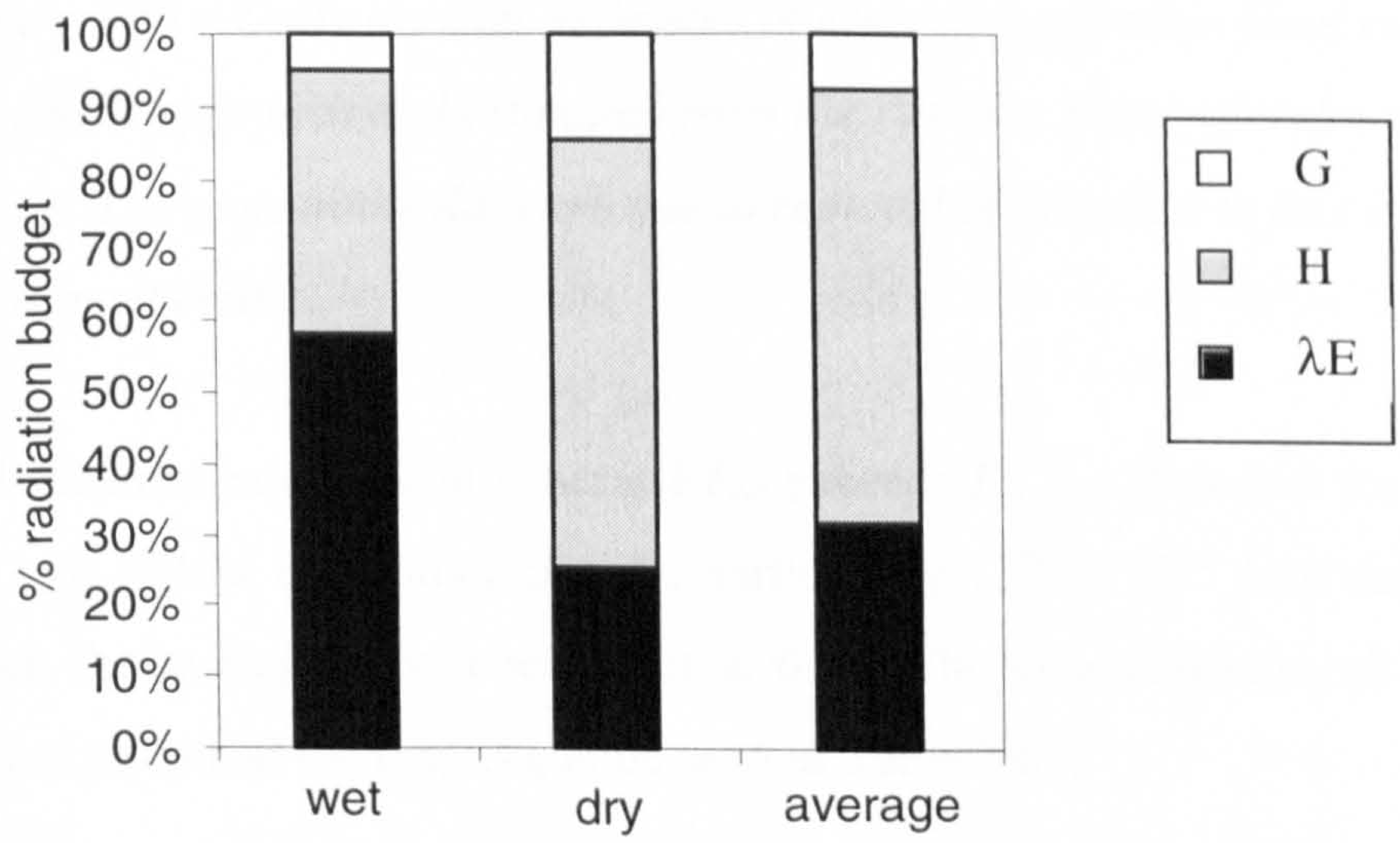


Figure 3.05 – Proportion of net radiation partitioned to each energy flux on a day with high radiation (19/6/01), a day with high rainfall (14/7/01) and the average for the period studied.

This again shows the importance of sensible heat in the energy balance at this site. Figure 3.05 also shows the energy balance for a typical wet day and a typical dry day. In general a higher proportion of energy was latent heat on wet days than on dry days. However the absolute flux of latent heat and hence evapotranspiration remained fairly constant whilst the sensible heat varies with the overall supply of radiation.

Figure 3.06 shows a plot of the evapotranspiration calculated using BREB and that calculated from the reference Penman Monteith equation.

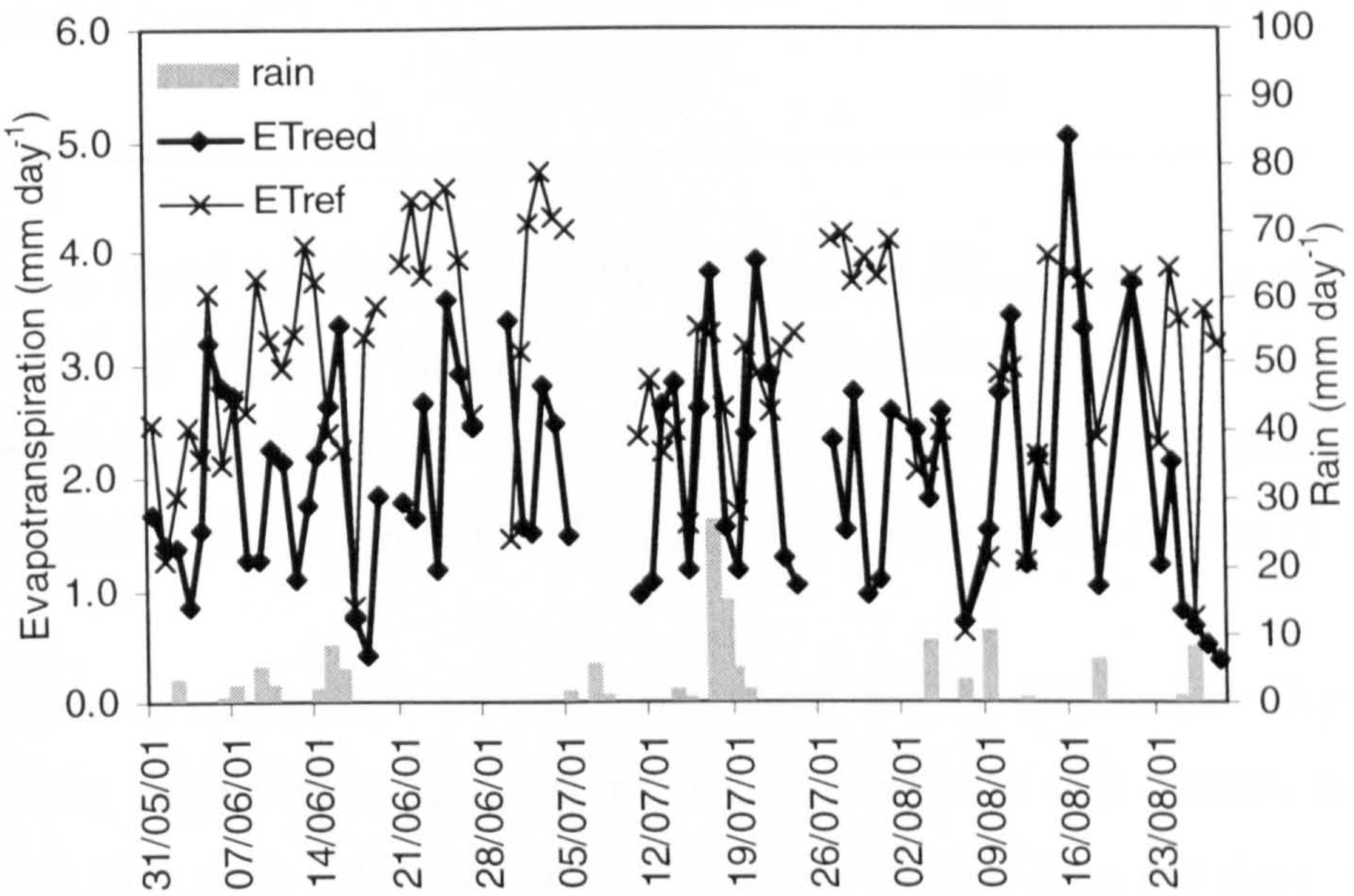


Figure 3.06 – Daily average estimates of evapotranspiration from reeds (ET_{reed}) using the BREB approach and estimated from the Penman Monteith reference equation (ET_{ref}). The gaps in the data are due to removal of data due to flux inconsistencies and instrument errors.

This demonstrates that in general ET_{ref} exceeds ET_{reed}. However for some short periods the two follow each other closely, particularly 13th to 22nd July and 3rd to 21st August when the average crop coefficient is 0.98. The characteristics of these periods were investigated and the results can be seen in Table 3.03.

Table 3.03 – Comparison of average daily rainfall and net radiation between periods for which ET_{ref} and ET_{reed} match up closely (13th-22nd July and 3rd-21st August) and those for which ET_{ref} is greater than ET_{reed} (all other dates) with P values indicating the significance of differences between the two groups.

	13th-22nd July and 3rd-21st August	All other dates	*P*
Kc	0.97	0.59	<0.001
Net radiation (W m^{-2})	220	265	0.014
Rainfall (mm d^{-1})	3.9	0.8	0.001
n	53	24	

It was found that the days on which ET_{ref} and ET_{reed} match up closely (13th-22nd July and 3rd-21st August) had significantly lower radiation and significantly higher rainfall and cloudiness. It appears that on days with a combination of greater cloud cover, low radiation and high rainfall the ET_{reed} is closer to ET_{ref} than it is on brighter, drier days.

It is to be expected that evapotranspiration will be different on days when it has been raining compared to days that are dry. It was found that overall, rain days (days with >= 0.1mm rain) tended to have a *Kc* closer to unity than dry days. The average *Kc* on days with rain is 0.88 as opposed to 0.53 on dry days (p=0.04). Although this is a significant difference, there is variation within both groups. Among the wet days this is partly related to the timing of rainfall – rain that falls at night will have much less impact on evapotranspiration than rain falling in the middle of the day. The high crop coefficients related to rainfall may only be present for some of the day when the canopy is actually wet. It must also be borne in mind that as temperature gradients are smaller on wet days, on average significantly less usable data were collected than on dry days (on average 40% of the data was unusable on wet days). However, on the days that were used it was considered that enough data were collected to be able to interpolate sensibly and have confidence in the results.

Daily rates of evapotranspiration (ET_{reed}) as measured by BREB plotted against that calculated for reference evapotranspiration (ET_{ref}) can be seen in Figure 3.07.

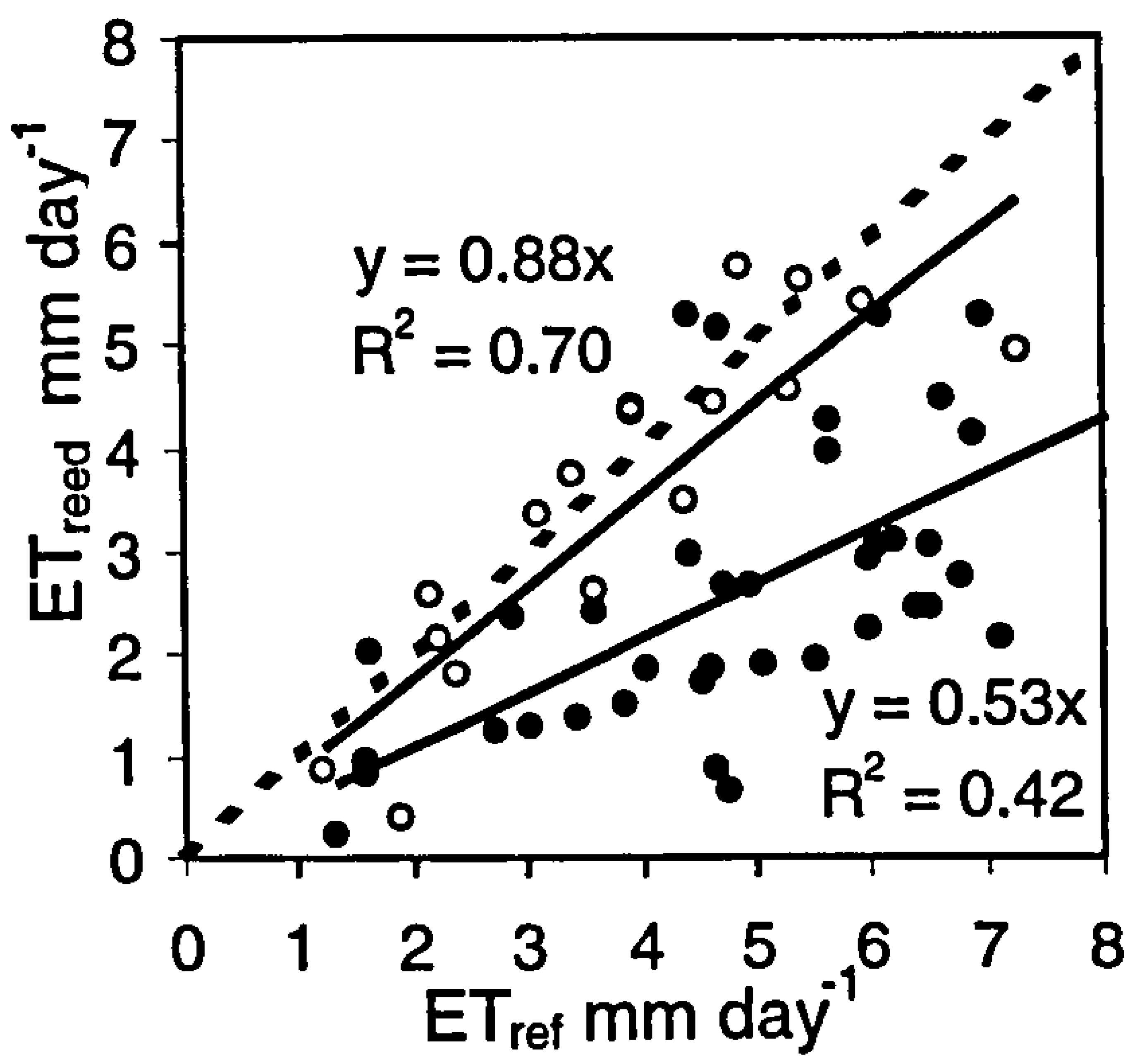


Figure 3.07 - The relationship between ET_{ref} and ET_{reed}. Open circles represent days with rain and closed circles represent dry days. Best fit lines forced through the origin are shown separately for wet and dry days (solid line) to create crop coefficients and the 1:1 line (dashed line) is also shown.

Although the overall best fit line (not shown in the figure) gives an average crop coefficient of 0.61, there is a very weak relationship ($r^2 = 0.34$) between ET_{ref} and ET_{reed} on a daily basis when all days were included. Plotting weekly averages did not improve the relationship. However when the days with and without rain are separated there is an improved relationship between ET_{ref} and ET_{reed}. This is particularly true for wet days ($r^2 = 0.70$) although there is still a lot of scatter among dry days ($r^2 = 0.42$). However both relationships are statistically significant.

3.5 DISCUSSION

The results must be considered in light of the limitations of the Bowen ratio approach and the assumptions which must be met for the data to be valid. It has generally been found that the turbulent transfer coefficients for vapour and heat are identical. Although this has been questioned in stable conditions (Lang *et al.* 1983; McNaughton and Laubach 1998), stability never occurred during the daytime measurement periods (see Appendix E). The second assumption is that the measurements of temperature and vapour pressure occur within the surface layer and there is no horizontal energy input. Measurements in this study were made within the fetch requirements for BREB and therefore advection should not be a large part of the energy balance. The roughness of the reed canopy compared with that of grass means that fetch requirements are reduced from the estimated 100 times the top measurement height to around 46 times. Measurements were made close to the top of the canopy within the roughness sublayer where flux-gradient relationships have been shown not to hold (Cellier and Brunet 1992). However the same authors show that although temperature and humidity profiles may be modified they are modified similarly so the first assumption of the Bowen ratio method is still valid at these heights.

If the assumption that evapotranspiration is zero at night is incorrect, this will lead to an underestimation of its daily value. There have been few measurements of ET_{reed} at night. Smid (1975) assumed zero evapotranspiration at night and Burba *et al.* (1999) found that fluxes of λE were small but variable at night – between –30 and +30 W m^{-2}. However Souch *et al.* (1996), using eddy correlation, found that evapotranspiration remained positive throughout the night sustained by a large energy release from storage. In the current study it was found that G was positive throughout the night, especially from inundated soils but this may be partially cancelled out by loss of long wave radiation from the canopy.

The energy balance that was created from the BREB data showed that the sensible heat component is significant and the evaporative potential of the net radiation is greater than actual conversion to latent heat. On the majority of days, measured evapotranspiration from reeds was less than reference evapotranspiration. The lack of a simple linear relationship between them may be anticipated in the light of the discrepancy in the physiological characteristics between reeds and the short grass-like crop that is the basis of reference evapotranspiration. The most obvious canopy difference is that of height and this leads to differences in surface roughness and atmospheric coupling – grass is poorly coupled to the atmosphere whereas reeds are well coupled so that the control of evapotranspiration from them is totally different (McNaughton and Jarvis 1983). Transpiration rates may also be controlled by physiological changes such as stomatal control and loss of turgidity. Reference evapotranspiration may therefore not be a useful method of modelling reed evapotranspiration. The results showed that there was a variation in crop coefficients dependent on meteorological conditions. On days with low radiation and cloud cover and particularly those with rainfall, i.e. days when the canopy was wet for at least some of the evaporative period, the crop coefficients were larger than on dry, bright days. This result was also found by Campbell and Williamson (1997). A starting point in using crop coefficients in reeds may be to use a different coefficient for days with rainfall and days without. This difference is possibly due to the fact that the rate of evaporation of intercepted raindrops is faster than the rate of transpiration from the plant, as it is not constrained by stomatal resistance. The resistance parameters within the Penman Monteith equation are also in reality not constant and stomatal resistance in particular is known to vary diurnally and with weather conditions and this may also account for some of the scatter.

The crop coefficient traditionally recommended for use in estimating reedbed evapotranspiration in the UK is 1.4 and higher results than this have been measured by Smid (1975) and Fermor *et al.* (2001) on the two smaller of their three sites. However, when applying crop coefficients with a grass reference under well watered conditions, for a vegetation expanse of greater than 50 m, the crop coefficients should not exceed 1.3 because there is insufficient energy available without an additional horizontal source (ASCE 1996). It is likely that these high values occurred on reedbeds when advection from surrounding dry lands was an additional energy source to the energy budget. In this study evapotranspiration was measured within a large reedbed and entirely within the boundary layer as shown by the satisfied fetch requirements. Large areas of homogenous vegetation often reduce evapotranspiration rates as the transpiration of the vegetation will raise the humidity near the surface, resulting in lower evapotranspiration and lower crop coefficients. In the UK large reedbeds are the exception rather than the rule (there are only eight reedbeds over 50 ha), so for general management purposes a crop coefficient must be selected that includes the likely advective influences of the site. However, it may be that on larger sites at least, a lower *Kc* than the standard 1.4 is required.

3.6 CONCLUSIONS

On the majority of days it was found that measured evapotranspiration rates from reeds were lower than reference evapotranspiration. However, crop coefficients varied with meteorological conditions – on low radiation cloudy days and particularly those with rainfall, i.e. days when the canopy was wet for at least some of the evaporative period, the evapotranspiration was much closer to reference as evapotranspiration of intercepted rainfall was not subject to stomatal control. Crop coefficients were lower than has often previously been assumed and measured and this can be partly attributed to the large area studied and lack of advective influences.

On a day-to-day basis, the crop coefficients were highly variable, indicating that the approach is not useful in creating models with high temporal resolution. If models of evapotranspiration on a daily time-step are required, a more successful approach may be to calibrate the Penman Monteith equation so that it can be used directly in a single step. This will require taking measurements of zero plane of displacement, roughness lengths and canopy stomatal conductance.

CHAPTER 4 – Estimating Evapotranspiration by Parameterising the Penman Monteith Equation Directly for Reeds

4.1 INTRODUCTION

4.1.1 Why use Penman Monteith?

The Penman Monteith equation is a useful and widely applied evapotranspiration model. It is based on physical assumptions and has proved satisfactory when compared to other methods (Jarvis *et al.* 1981). Its combination approach includes both aerodynamic and energy balance terms and it has achievable data demands requiring meteorological data at only one level (Thom and Oliver 1977). "Calculations based on the Penman Monteith equation probably provide the most powerful general technique for estimating evapotranspiration rates from vegetation" (Jones 1992 p.124). The Penman Monteith equation is:

$$\lambda E = \frac{s(R_n - G) + \rho CpD / r_a}{s + \gamma(1 + r_s / r_a)} \qquad (4.01)$$

where s is slope of the saturation vapour pressure curve (kPa °C^{-1}), R_n is net radiation (W m^{-2}), G is ground heat flux (W m^{-2}), ρ is mean air density (kg m^{-3}), C_p is specific heat of air at a constant pressure (1.13 kJ kg^{-1} °C^{-1}), D is the vapour pressure deficit at reference height (kPa), γ is the psychrometric constant (kPa °C^{-1}), r_a is aerodynamic resistance (s m^{-1}) and r_s is surface resistance (s m^{-1}).

The FAO Irrigation and Drainage Paper 24 (Doorenbos and Pruitt 1977) recommended the use of Penman-type equations together with crop coefficients to estimate crop evapotranspiration. However the FAO Penman methodologies were found to have weaknesses and subsequently a hypothetical reference crop described by the Penman Monteith equation was substituted for the original living reference crop (Allen *et al.* 1994c). Using this technique the Penman Monteith equation is used to calculate crop evapotranspiration (*ET_c*), which is defined as evapotranspiration from a specified vegetative surface in good agronomic condition, extensive in area, under optimum soil water conditions and achieving full production under the given climatic conditions. ET_c varies according to meteorological variables – primarily radiation, temperature, vapour pressure deficit and windspeed – and also according to canopy characteristics including leaf surface area, the plant height and roughness, reflection and ground cover of the leaves. This second group of factors change with season and growth stage and are therefore difficult to quantify. The reference surface defines these features by setting values for the canopy parameters in the Penman Monteith equation. These parameters are similar to those for a grass surface. The crop coefficient (Kc) is intended to represent an integration of the effects of the canopy characteristics that distinguish the crop from the reference surface (Allen *et al.* 1998). However using a single coefficient over a period of months or over a growth stage can prove inadequate as these factors can vary on shorter time-scales.

4.1.2 Using Penman Monteith in one step

In recent years there has been renewed interest in using the Penman Monteith equation in a one-step approach without the use of crop coefficients (Alves and Pereira 2000, Herbst *et al.* 1996) by parameterising it for a particular crop. Following this technique for reeds will allow its use in directly modelling evapotranspiration, whilst simultaneously providing information regarding the physical and physiological processes governing the evapotranspiration flux from reeds. It may also help explain the results that were measured by the Bowen ratio approach in the previous chapter, including the fact that measured evapotranspiration was found to be lower than has previously been assumed. In addition to the relatively simply measured meteorological parameters, the Penman Monteith equation also contains two parameters which describe physiological factors which must be determined specifically for reeds. These are surface resistance (r_s) (or its inverse surface conductance (g_s)) and aerodynamic resistance (r_a) (or its inverse aerodynamic conductance (g_a)). It is a lack of a methodology for determining these parameters that restricts wider use of the Penman Monteith equation in one step to estimate evaporation.

4.1.3 Assumptions of the Penman Monteith equation

The Penman Monteith equation assumes that the pathways of heat, momentum and water transfer are similar and that the canopy can be treated as a single big leaf. This has caused controversy when applied to the canopy, as there are obvious inaccuracies introduced in this simplification, especially in sparse canopies. However comparisons with other models show that the errors are not serious. In principle the problem can be avoided by applying the model separately to several layers of the canopy. However this requires separate data for each layer (Jarvis *et al.* 1981). The Penman Monteith equation also assumes a steady state environment and negligible cuticular transpiration and soil evaporation.

4.1.4 Aerodynamic resistance

Aerodynamic resistance may be calculated from the following equation (Allen *et al.* 1994c):

$$r_a = \frac{\ln[(z-d)/z_{oh}]\ln[(z-d)/z_o]}{k^2 u} \qquad (4.02)$$

where u is windspeed (m s^{-1}), k is Von Karman's constant (0.41), z_o is roughness length governing momentum transfer (m), d is height of zero plane of displacement (m), z is height of windspeed measurement (m) and z_{oh} is roughness length governing heat and vapour transfer (m). The parameters z_o (momentum roughness length) and d (zero plane of displacement) must therefore be determined. The value of z_o is dependant on the aerodynamic roughness of the surface and a rough surface such as reeds will have a larger value than that which would be specified for a smooth grass surface. The zero plane of displacement is the mean level at which momentum is absorbed by the plant community.

4.1.5 Surface resistance

Determination of surface resistance is the biggest problem in using the Penman Monteith equation as it is the hardest of all canopy variables to measure (Rochette *et al.* 1991, Jarvis *et al.*1981). The value of r_s is influenced by physiological, climatic and edaphic factors – temperature, radiation, water stress, position of leaves in the plant and the age of the leaves (Alves and Pereira 2000). Stomatal resistance at the leaf and canopy scale have been correlated with many environmental conditions particularly radiation, vapour pressure deficit, temperature and aerodynamic resistance. Low levels of radiation cause an increase in resistance at the beginning and end of the day and in the lower leaves in the canopy. Resistance decreases as temperature rises, up to a maximum point, then increases at high temperatures due to leaf water deficits or large leaf to air water vapour concentration differences and increases with increasing relative humidity due to a reduction in the concentration gradient (Turner 1991). Soil water potential, which significantly influences transpiration in some environments, is less important in wetlands where soils are permanently waterlogged.

Three methods have been used to determine the surface resistance of vegetation. The first is the "bottom up" approach which involves measuring stomatal conductance at the leaf level and scaling up to canopy level (e.g. Lafleur 1988; Takagi *et al.* 1998). The second method is the "top down" approach which involves rearranging the Penman Monteith equation where latent energy is known, so that surface resistance can be found (e.g. Baldocchi *et al.* 1991; Alves *et al.* 1996). Thirdly stomatal resistance can be modelled from meteorological parameters (e.g. Jarvis 1976; Stewart 1988).

4.1.5.1 Bottom up approach

Stomatal behaviour has been described as the key to the apportioning of λE and H (Lhomme 1998) and in many cases transpiration seems to be controlled by stomata rather than aerodynamic conditions (Sanchez-Carrillo *et al.* 2001). Even with unlimited soil moisture, surface resistance is never negligible but has a significant residual value (unless the surface is actually wet) (Thom and Oliver 1977). Stomatal conductance and transpiration characteristics of wetland plants have not been studied extensively and little is known about the response of their stomata to environmental variables (Takagi *et al.* 1998).

Conductance of individual leaves can be measured directly as long as a suitable sampling procedure is employed. Porometers are the most common method of measuring transpiration, and have been used in several studies in wetlands (e.g. Jones and Muthin 1984; Lafleur 1988; Abtew *et al.* 1995; Takagi *et al.* 1998). In porometers with an open system, a flow of air passes through the chamber and over the leaf. Transpiration rate is calculated from the difference in water vapour content of air entering and leaving the cell, or from the time period it takes for air entering the system to raise the moisture content of the air leaving the system to a certain level. Many observations are required, though fewer are needed in an annual crop where leaf age is less important. Turner (1991) suggests 15 randomly sampled measurements on both sides of the leaf in the upper middle and lower layers of the canopy should be sufficient to estimate stomatal conductance at a particular time. Ideally a uniformly sunny or uniformly cloudy day is required. It is necessary to have sufficient measurements to account for the partial closing of stomata due to shading by overlying plant cover (Munro 1989).

A more accurate technique is to use an Infra Red Gas Analyser (IRGA) to measure the concentration of water vapour within a cuvette held around the leaf. Heteroatomic gases such as H_2O absorb infra red radiation in very distinct wavebands. An IRGA measures the density of a particular gas present in a carrier by the reduction in transmission of infra red radiation at one or more of these wavebands (Jarvis and Sandford 1985). Although this technology has been used in the measurement of CO_2 concentration since the late 1950s, it is only recently that several portable infra red gas exchange systems have been developed for use in the field. Field instruments need to be tolerant of vibrations and wide fluctuations in temperature and humidity and also lightweight and operable from batteries. Infra red gas analysers are typically used to measure CO_2 but they can also measure water vapour. These are designed to measure net photosynthesis and transpiration rates as well as stomatal conductance. This is one of the most accurate ways of measuring transpiration (Pearcy *et al.* 1989). Examples of the use of an IRGA to measure stomatal conductance include the work of Vanyarkho (1996) and Smith *et al.* (1994).

The most difficult aspect of the bottom up approach is scaling up from leaf to canopy level. Particular problems have been found when measuring the resistance of complete cover crops or over wet soil (Alves and Pereira 2000). This is because in complete cover crops not all the leaves contribute equally to transpiration and there are humidity gradients within the canopy. Porometers or IRGA systems usually only enclose part of the leaf and there is variability within the leaf, among leaves in a canopy layer and between leaves. Stomatal resistance increases down into the canopy as the light decreases and the bottom of the canopy often contributes little to r_s (Saugier and Katerji

1991). In order to deal with this problem the concept of effective leaf area was introduced and the leaf area index multiplied by 0.5 to account for this. Equation 4.03 is the most commonly used method for scaling up.

$$r_s = \frac{r_l}{0.5LAI} \qquad \text{(Allen } et\ al.\ \text{1989)} \qquad (4.03)$$

where r_l is the average daytime value of stomatal resistance for a single leaf. Some have argued that scaling up is not possible at all because the assumption that heat, water and momentum have sink and source at the same level is unrealistic (Saugier and Katerji 1991).

4.1.5.2 Top down approach

In its original form, the Penman Monteith equation set out not to be a means to calculate evapotranspiration but to estimate surface resistance. Where λE is known it is possible to determine r_s from an alternative arrangement of the Penman Monteith equation. This is an attractive approach as it is based on simple and general laws and has few, simply measured input variables (Baldocchi *et al.* 1991). The technique is less informative about the components making up r_s than the bottom up approach but it is easier to create a single value for canopy resistance (Jarvis *et al.* 1981).

The bottom up and top down approaches would not necessarily be expected to produce identical results, although they are often used interchangeably. Monteith (1965) expected r_s to be connected with stomatal resistance and fundamentally a physiological parameter. In a well coupled environment they may not differ significantly (Thom 1975) and some authors have presented results which agree well (e.g. Grantz and Meinzer 1991; Herbst 1995). However the top down r_s contains additional non-physiological information pertaining to the net radiation balance and aerodynamic conductance in the canopy and also includes soil evaporation.

4.1.5.3 Modelling surface resistance from meteorological variables

Measurements of stomatal resistance in the field are laborious and can only be done when the foliage is dry. Because of this, there has been a large amount of work attempting to predict surface resistance from meteorological variables. Cain (1998) gives a useful review of the literature. Models may be based on direct studies of measured stomatal conductances of leaves either in controlled or natural environments (e.g. Jarvis 1976) or calibrated from estimates of r_s obtained from the top down approach (e.g. Stewart 1988).

Several types of model have been tried. The simplest is multiple regression analysis between r_s and influential meteorological variables but as this is entirely empirical it is generally poor at predicting the complex responses that are due to the non-linear response of r_s to most meteorological variables. Boundary line analysis can also be done where a line is drawn around the outside of the data mass but this requires a spread of data across the whole range of possibility which is rarely available. The most commonly used and recommended model in the literature is that first proposed by Jarvis (1976). Examples of its use include Baldocchi *et al.* (1991) and Stewart (1988). It was originally expressed in terms of stomatal conductance but for consistency within this chapter it is expressed here as a resistance following Lhomme (1998). Stomatal resistance is expressed by a minimum resistance multiplied by a series of independent stress functions, each representing one variable. The general Jarvis model is:

$$r_s = r_{s\,\min} f(Rs) f(T) f(D) f(\psi) \qquad (4.04)$$

where $r_{s\min}$ is minimum stomatal resistance in optimal conditions, Rs is solar radiation, T is air temperature, D is vapour pressure deficit and ψ is leaf water potential. The model is based on the presumption that the form of the model can be determined in the laboratory and parameters determined by fitting the model to field data using non-linear least squares regression and that r_s responds to variables independently. The usual inputs are photon flux density at the leaf, air temperature and vapour pressure deficit above the canopy and either a set of measurements of r_s to find the parameters or a set of parameters to predict r_s (Jarvis *et al.* 1981).

Each function varies from one to infinity and a large number of functions have been determined for each parameter. The original model was fitted to forest data. It was found that most of the variation was accounted for by photon flux density and vapour pressure deficit (r^2 >0.7) (Jarvis *et al.* 1981). However the model has disadvantages in that there is an assumption that there are no interactions and the parameters, once defined have little physiological meaning because the model is descriptive rather than mechanistic.

Other types of model have also been employed. Herbst (1995) used a model with just photon flux density and vapour pressure deficit that explained 70% of the variation. More mechanistic models have also been developed such as those of Collatz *et al.* (1991). These models attempt to include the often observed correlation between photosynthesis and r_s and have achieved good results (r^2 = 0.92). However these models require large data inputs including photosynthetic rate so they are not very easy to use in practice.

A general problem is that the models are specific to particular vegetation and climate systems and parameters must be readjusted for each new situation. The complexity of a model is a balance – lots of complexity is required to achieve realism but considerable simplification is required if the model is to be useful and allow predictions of actual evapotranspiration (Raupach and Finnigan 1988).

4.1.6 Atmospheric coupling

The roughness of the vegetation surface is important in the estimation of evapotranspiration as it determines the degree to which the atmosphere is coupled to the canopy. When r_a is small, heat and mass transfer are very efficient and there are small gradients through the boundary layer so that leaf temperature and air temperature are similar and independent of R_n. The leaf is then well coupled to the atmosphere. Vapour and heat fluxes from the leaf do not cause a change in the saturation deficit of the leaf surface. The conditions of the atmosphere are imposed on the leaf ("imposed evaporation") and evapotranspiration is proportional to stomatal resistance. At the other extreme when r_a is large and the heat and mass transfer is small, the vapour pressure deficit at the leaf reaches a local equilibrium ("equilibrium evaporation"). The surface is poorly coupled to the atmosphere and evapotranspiration is much less dependent on stomatal resistance. McNaughton and Jarvis (1983) introduced a factor to quantify atmospheric coupling (Ω) which gives an indication of a canopy's sensitivity to stomatal resistance. It is a ratio between equilibrium evaporation and imposed evaporation. The atmospheric coupling factor gives us an indication of the canopy's sensitivity to stomatal resistance.

4.1.7 The influence of a wet canopy

Surface resistance is analogous to stomatal resistance in dry conditions but is zero when the canopy is completely wet. If it is partially wet it should be between 0 and the dry value of r_s (Stewart 1977). Transpiration is reduced by the presence of intercepted water but evapotranspiration occurs at substantially higher rates than from dry leaves as the evaporation of intercepted water can be up to four times the rate of transpiration (Stewart 1977).

4.2 OBJECTIVE

The objective of this chapter is to parameterise the Penman Monteith equation for reeds in order to estimate evapotranspiration directly. This data will be compared with the crop coefficient technique and allow investigation of reed evapotranspiration more closely in order to discover what factors it is influenced by. This may help explain the values of evapotranspiration measured in Chapter 3, in particular the factors which constrain evapotranspiration below those of reference grass.

4.3 MATERIALS AND METHODS

4.3.1 Study site and period

The data was collected on Stodmarsh National Nature Reserve. The majority of the data analysed in this chapter is the same as that described in Chapter 3 where the experimental set-up is fully described. Data was collected between 30/05/01 and 31/08/01, and 30/04/02 and 16/07/02.

4.3.2 Measurement of meteorological parameters

Net radiation was measured directly using a net radiometer (Kipp and Zonen NR-lite) just above the maximum height of the reed canopy. Ground heat flux was measured using a combination of soil heat flux plates and thermocouples. The methodology used is discussed further in Chapter 3 and Appendices A and B.

Calculations were made following Allen *et al.* 1994b. The slope of the saturation vapour pressure curve was calculated from:

$$s = \frac{2504\exp\left(\frac{17.27T}{T+237.3}\right)}{(T+237.3)^2}$$ (Tetens 1930 as cited in Allen *et al.* 1994b) (4.05)

where T is air temperature. Air temperature was taken from the lower temperature measurement of the Bowen ratio equipment. Atmospheric density was calculated from:

$$\rho = \frac{1000P}{T_{kv}R} \qquad (4.06)$$

where P is estimated atmospheric pressure (kPa) and R is the specific gas constant (287 J kg^{-1} K^{-1}).

$$T_{kv} = T_k\left(1 - 0.378\frac{e_d}{P}\right)^{-1} \qquad (4.07)$$

where T_k is temperature (K) and e_d is vapour pressure at dewpoint (kPa)

Saturation vapour pressure is:

$$e_a = 0.611\exp\left(\frac{17.27T}{T+237.3}\right) \qquad (4.08)$$

Vapour pressure at dewpoint was also calculated from Equation 4.08 substituting the dewpoint temperature (°C) as measured by the dewpoint hygrometer for air temperature. The psychrometric constant is:

$$\gamma = \frac{C_p P}{\varepsilon\lambda} \times 10^{-3} \qquad (4.09)$$

where ε is the ratio of molecular weight of water vapour to dry air (0.622) and λ is the latent heat of vaporisation (MJ kg^{-1}).

4.3.3 Determination of aerodynamic resistance

4.3.3.1 Calculating d and z_o

In order to determine r_a using Equation 4.02, z_o, z_{oh} and d were derived using wind profiles. The logarithmic wind profile equation is:

$$u = (u*/k)\ln(z - d/z_o) \qquad \text{(Thom 1975)} \qquad (4.10)$$

where $u*$ is friction velocity (m s^{-1}).

A wind profile requires a minimum of three anemometers at different heights. These were calibrated against a standard using an air conditioning unit. A second additional pole (attached above the extra pole described in Chapter 3) was added to the top of the tripod to create a suitable height range for the profile (Figure 4.01). Anemometers were connected to the datalogger and windspeed was averaged over five minute periods in 2001 and 10 minute periods in 2002 (5 minute periods were found to create too much data for the datalogger memory).

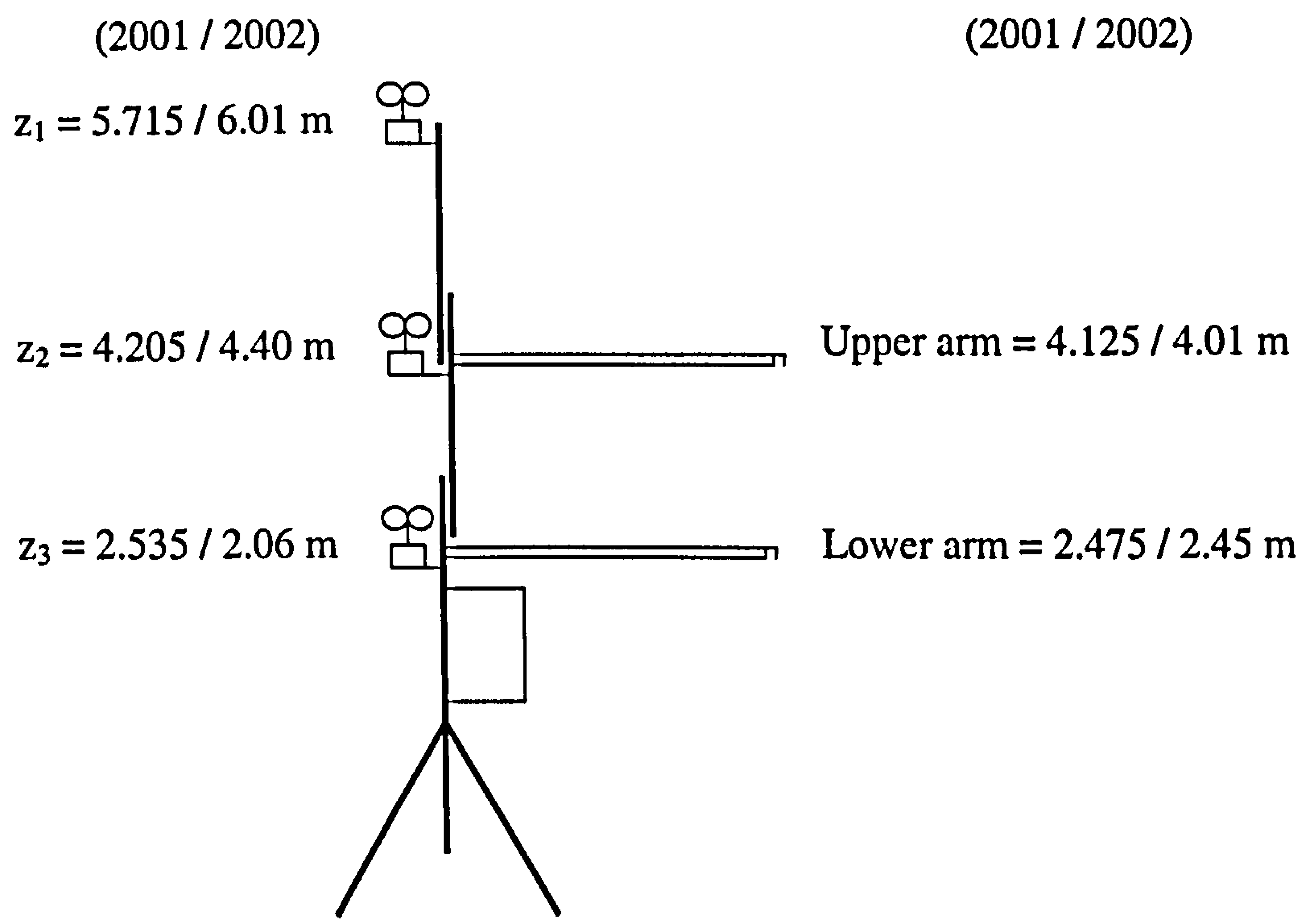


Figure 4.01 – Diagram showing the position and heights of instruments on the tripod with the heights in 2001 followed by those in 2002.

In order to work out d, Equation 4.10 was rearranged in the form of Equation 4.11. This required the values of windspeed (u_1, u_2, u_3) at the three heights (z_1, z_2, z_3) and the use of iteration to solve for d:

$$\frac{u_1 - u_2}{u_1 - u_3} = \frac{\ln(z_1 - d) - \ln(z_2 - d)}{\ln(z_1 - d) - \ln(z_3 - d)} \qquad (4.11)$$

Not all the data sets measured could be used. In some cases it was impossible to calculate a reasonable value of d and the result was a negative number. These data were excluded. The values of ln(z-d) were plotted against the windspeed in a straight line relationship (an example data set is shown in Figure 4.02) and z_o was calculated as the inverse log of the intercept of the line with the y axis, as z_o is equal to z-d when windspeed is zero. These calculations were performed for each data set using linear regression to calculate the intercept of the line.

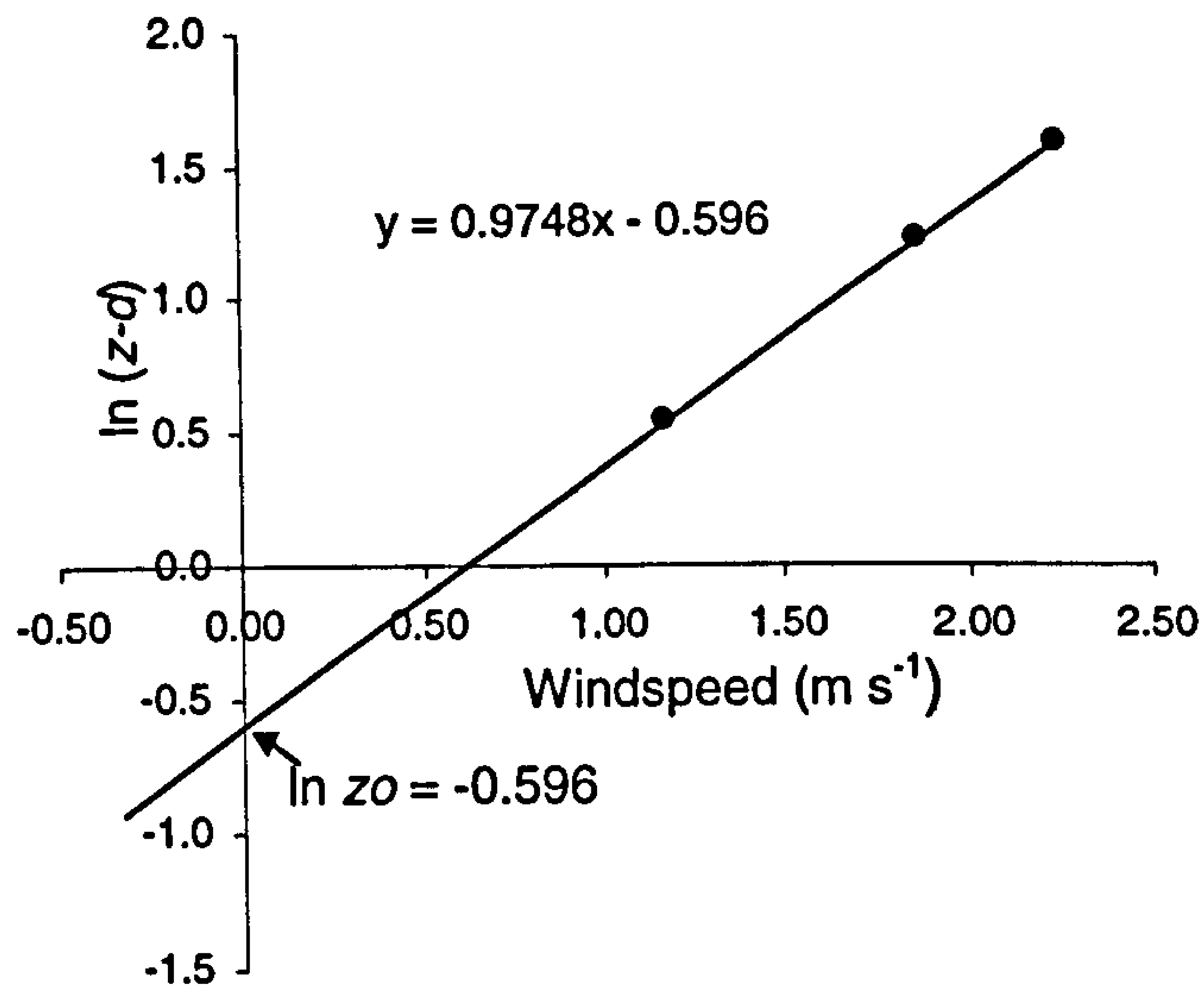


Figure 4.02 – An example of the estimation of z_o using a straight line plot of u against ln(z-d).

Because of the enhancement of momentum transfer by the pressure drag of a rough surface and the effects of bluff body transfer which affect z_o and not z_{oh}, the value of z_{oh} is smaller than z_o and was calculated as 0.1 z_o (Brutsaert 1975).

4.3.3.2 Crop Height

Aerodynamic resistance is linked to the height of the vegetation stand. Therefore the height of the reeds was measured on a weekly basis during the measurement period in 2002 and on a single date in August 2002 using a tape measure. On each occasion 10-20 random stems were chosen and measured and the maximum and mean height calculated.

4.3.4 Determination of surface resistance

Leaf area index *(LAI)* and the inverse of the measurement of leaf stomatal conductance were used to calculate r_s using Equation 4.03. The methodologies for measuring stomatal conductance and leaf area index are described below.

4.3.4.1 Bottom up approach – measuring stomatal resistance

In order to measure the stomatal resistance of individual leaves a combined infra red gas analyser (IRGA) system (PP Systems Ciras-1) was used. It measures both CO_2 and H_2O simultaneously, employing a differential open system. CO_2 and H_2O are measured together because they are interdependent, using a largely shared diffusion pathway and the stomatal responses that decrease H_2O movement also decrease CO_2. The differential system calculates photosynthesis from the CO_2 depletion that occurs as the air flows at a known rate past a photosynthesising leaf. The instrument employs four infra red gas analyser systems to measure the concentration of H_2O and CO_2. Infra red light is shone through a gas sample and the amount absorbed is calculated. Flow through the instrument is also measured (Field *et al.* 1989).

Sampling was done over four days (29/08/01, 28/05/02, 03/06/02 and 11/07/02) from the time when the dew had dried on the leaves in the morning until early evening. Some data was lost when it began to rain as measurements could not be taken whilst the leaves were wet. On each day 30 leaves were sampled per hour and the measurements were repeated on the same leaves each hour. The hour was divided into three twenty-minute periods in order to coincide with the measurements taken with the Bowen ratio equipment. In each twenty-minute period a separate set of 10 leaves were sampled, 3 from the top of the canopy, 4 from the middle and 3 from the bottom following the techniques of Rochette *et al.* (1991) and Herbst *et al.* (1996). If a leaf was damaged during sampling another leaf was chosen at a similar height with a similar degree of shading.

4.3.4.2 Leaf area index measurement

LAI was measured using a Canopy Analysis System (Sunscan, Delta–T Devices) (Figure 4.03). The canopy analysis system is a portable instrument that measures light levels of photosynthetically active radiation and directly calculates LAI.

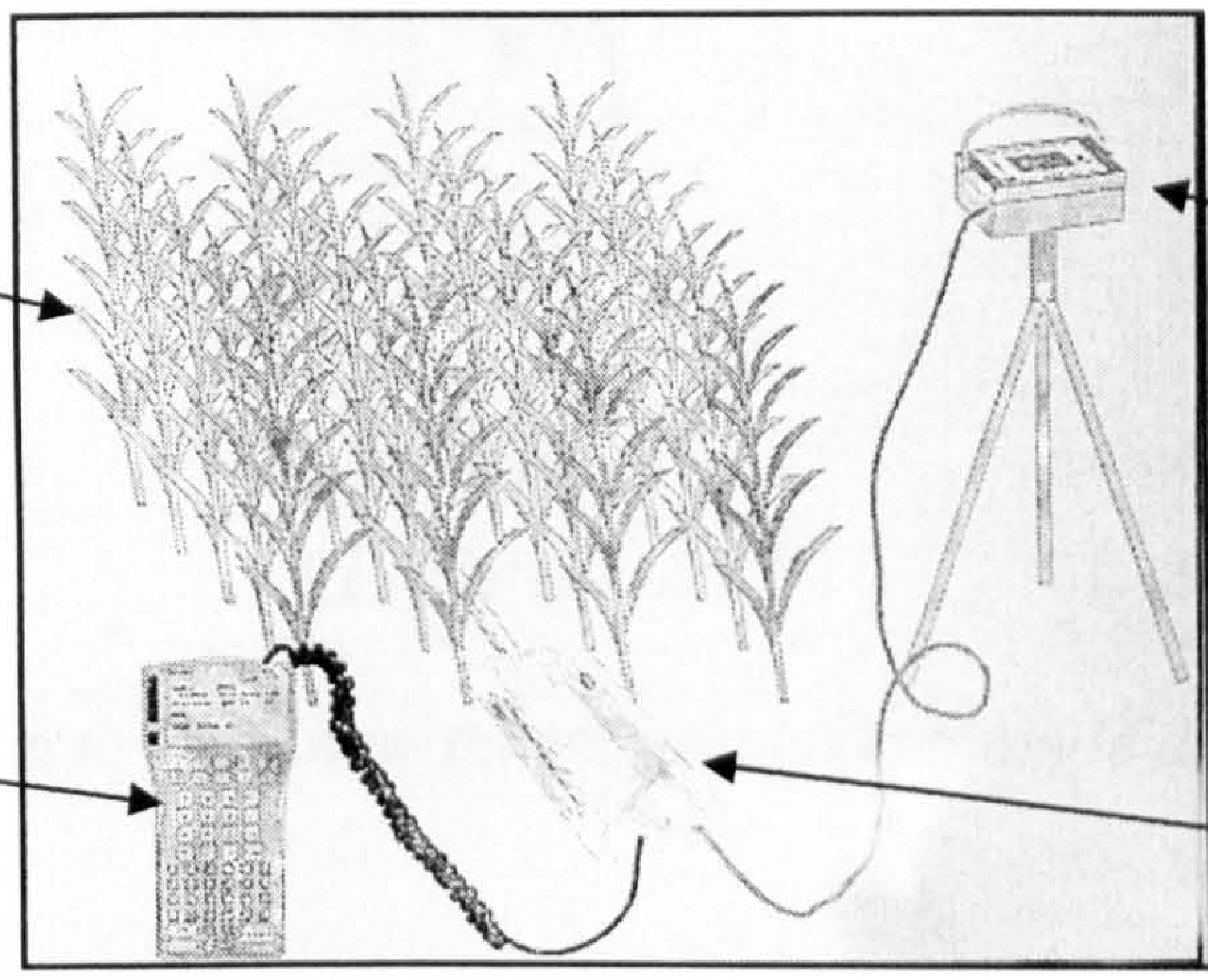
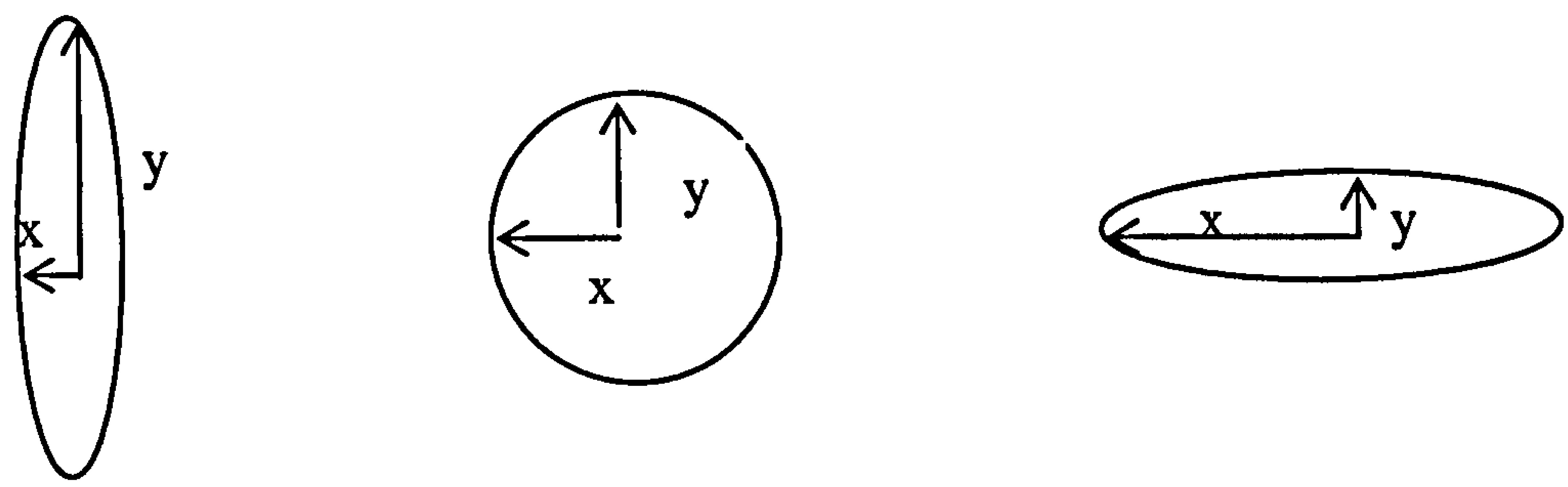


Figure 4.03 –Diagram of the set up of the Delta-T Sunscan Canopy analysis system

The beam fraction sensor was placed outside the canopy in order to measure light levels incident upon the canopy. This was placed close to the canopy whilst avoiding any shading by it. It has two photodiodes, one of which is shaded from the direct solar beam, in order to measure separately both the direct and diffuse light components. To calculate *LAI* the sunscan also requires information on the distribution of leaf angles, absorption of light within the canopy and knowledge of the solar (zenith) angle calculated from the time of day, the latitude and the longitude.

Most leaves have absorption values of 0.8-0.9 and as the leaves of *Phragmites communis* are not particularly dark or transparent a value of 0.85 was considered adequate as this is generally thought to be suitable for most leaf types (Potter *et al.* 1996).

The Ellipsoidal Leaf Angle Distribution Parameter (ELADP) was introduced by Campbell (1986) and modified by Wang and Jarvis (1988). It characterises the horizontal or vertical tendencies of leaves in a canopy in a single parameter. It is assumed that the canopy is distributed in space in the same directions as an ellipsoidal shape and the ELADP is the ratio of the horizontal to vertical axis of this shape (Figure 4.04)



Tall thin ellipsoid
ELADP low (min 0)

Spherical:
ELADP = 1

Broad flat ellipsoid:
ELADP high (max 1024)

Figure 4.04 – Diagram to show how the ellipsoidal leaf angle distribution parameter is calculated

In order to estimate the ELADP of the reed canopy, 20 stems were randomly selected. For each stem, each leaf was designated as being greater than 45° from vertical (mostly horizontal – *Nh*) or less than 45° from vertical (mostly vertical – *Nv*). Where leaves were bent the angle at the widest point in the leaf was used.

$$ELADP = \frac{\pi Nh}{2Nv} \tag{4.12}$$

The $\pi/2$ factor is related to the fact that the leaves are distributed all around the vertical axis so some are seen face on by a beam of light and some edge on.

Due to limitations caused by the length of the cable and difficulties of getting cable through the reeds, measurements were necessarily made within a few metres of the edge of the reedbed. However due to the density of the reed canopy this was thought to be representative. Readings were taken at a high and low level in the canopy.

4.3.4.3 Top down approach

This involves calculating r_s from a re-arrangement of the Penman Monteith equation:

$$r_s = r_a\left(\frac{s}{\gamma}\beta - 1\right) + \frac{\rho CpD}{\gamma\lambda E} \tag{4.13}$$

λE and β were found using the Bowen ratio energy balance technique and all other parameters were calculated as above.

4.3.5 Atmospheric coupling

Coupling is an expression of the sensitivity of transpiration to a change in stomatal or canopy conductance. It was calculated using the following equation:

$$\Omega = \frac{(s/\lambda)+1}{(s/\lambda)+1+(g_a/g_c)} \qquad \text{(Jarvis and McNaughton 1986)} \qquad (4.14)$$

The coupling factor (Ω) was calculated using the data from July 2001. Measured values of z_o and d were used to define g_a. The inverse of top down surface resistance values were used to create the required canopy conductance values.

4.4 RESULTS

The following sections describe the results of the aerodynamic resistance measurements – windspeed measurements and the calculated values of d and z_o. These are then related to crop height measurements. Surface resistance results are then presented beginning with mean daily top down results, and then the LAI and IRGA measurements used to create the bottom up estimations of surface resistance. Top down and bottom up stomatal resistance are then compared before a number of attempts at modelling stomatal resistance are presented.

4.4.1 Aerodynamic resistance

The mean windspeed measurements at the three heights for 2001 show the characteristic logarithmic profile that would be expected (Figure 4.05).

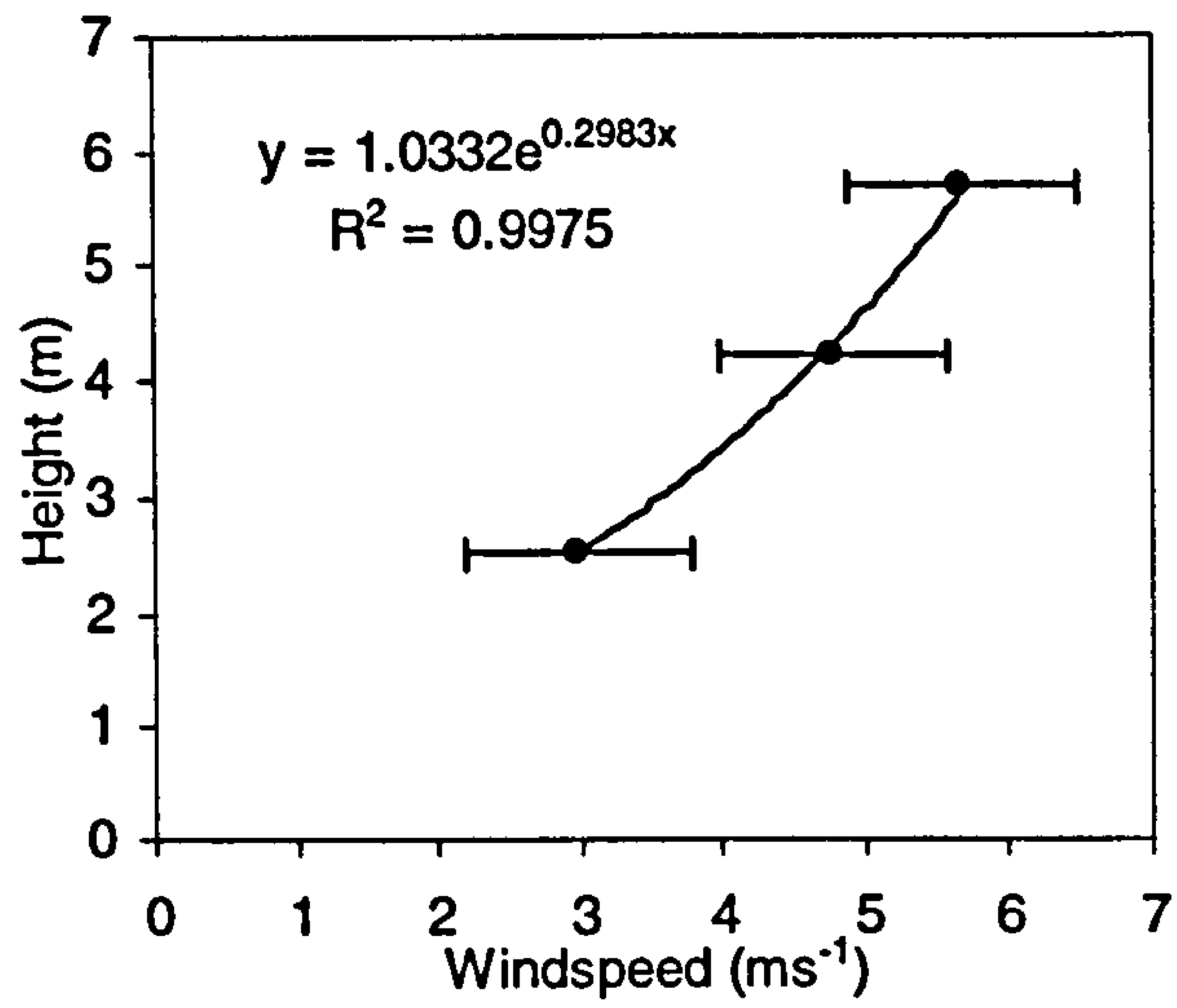


Figure 4.05 – Mean windspeed measurements (2001) at three heights with standard error bars. n = 7449

The values for d and z_o were variable between each measurement period. The mean value for d was consistent between 2001 and 2002 but z_o was smaller in 2002 than 2001 (Table 4.01).

Table 4.01 – Mean seasonal values for d and z_o with their standard errors

	d (m)		z_o (m)		n
	Mean	S.E.	Mean	S.E.	
2001	1.005	0.024	0.48	0.015	1826
2002	1.03	0.015	0.32	0.007	1187

With a typical windspeed of 4.0 m s^{-1}, the average values for 2002 result in a value of r_a of 6.52 s m^{-1}. This is compared to an aerodynamic resistance of 56.06 s m^{-1} using values of d and z_o for the reference (grass) surface and the same windspeed.

4.4.1.1 Vegetation height

Vegetation growth was rapid through May and June and reduced in rate in late June and July until maximum crop height was achieved by the beginning of August (Figure 4.06).

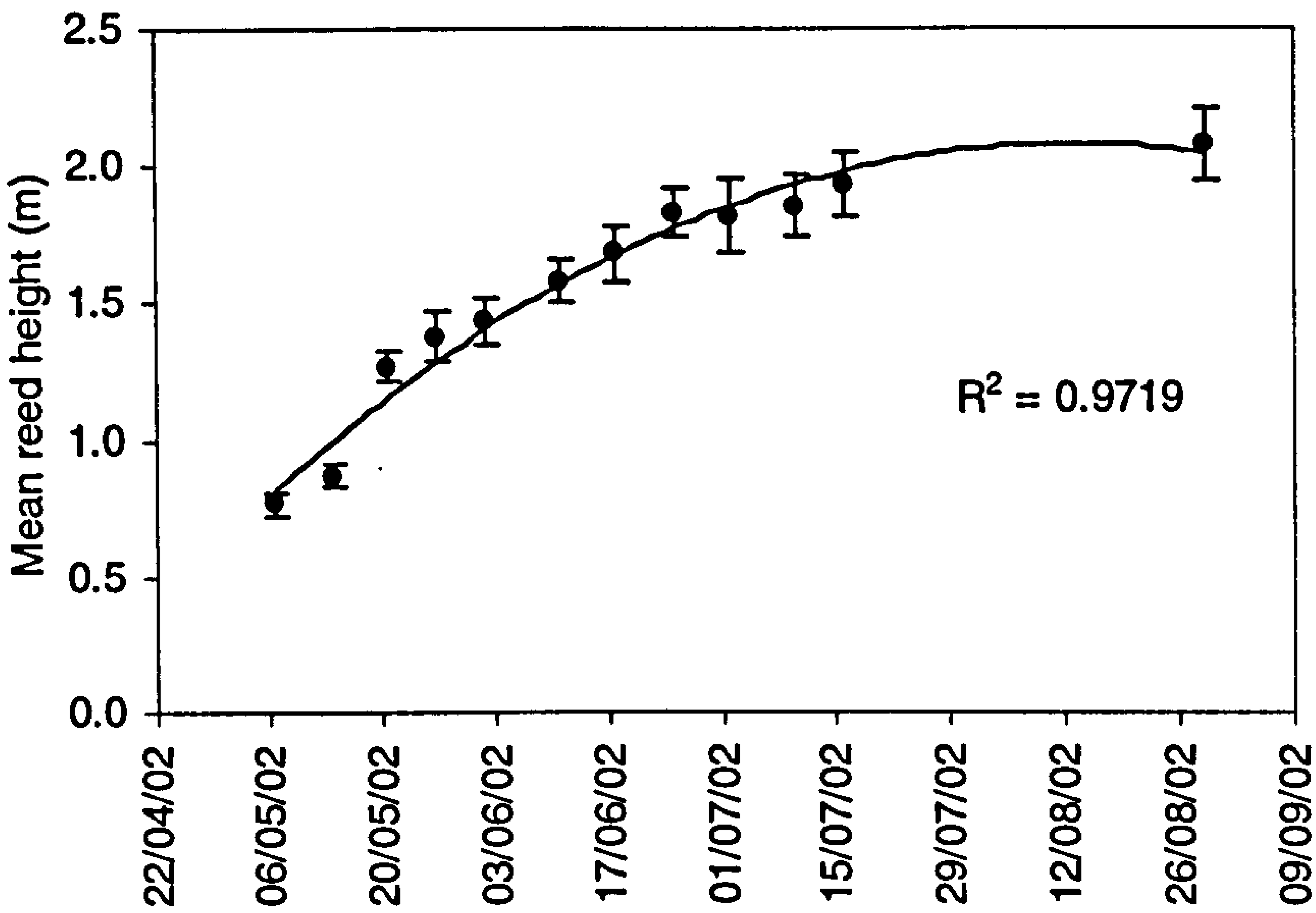


Figure 4.06 – Height measurements in 2002 with 95% confidence intervals and the best fit line.

4.4.1.2 Relationship between crop height and d and z_o

There is a theoretical relationship between the canopy height and *d* and z_o. However there appeared to be no systematic change in these parameters as the crop height increased (Figure 4.07). This indicates that a single average value may be used throughout the season.

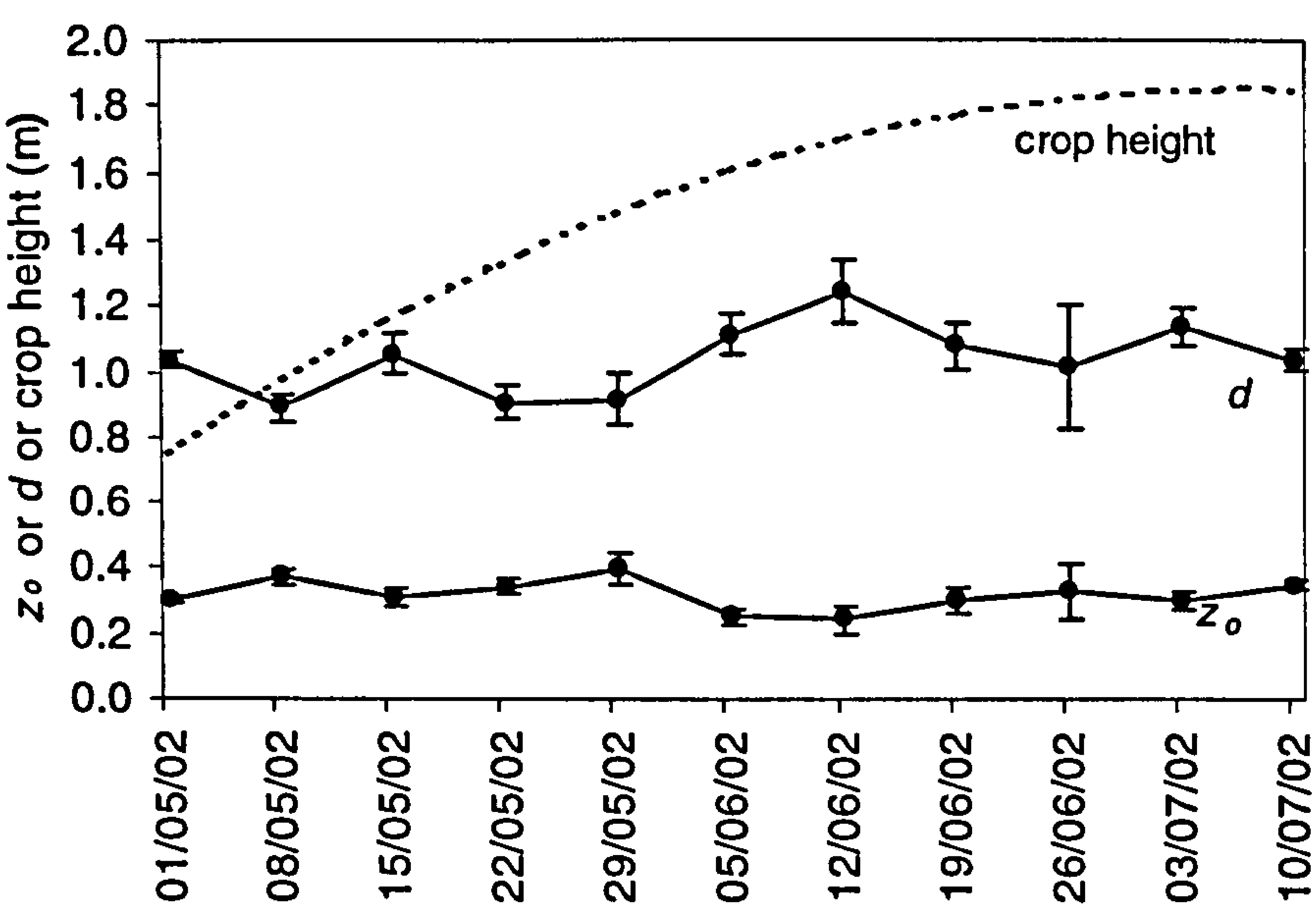


Figure 4.07 – Weekly means of d and z_o with 95% confidence intervals (solid lines) compared to the best fit line of the weekly vegetation height measurements (dashed line).

4.4.2 Determination of surface resistance – top down approach

Top down stomatal resistance was calculated using mean measured *d* and z_o. There is a lot of variation in the mean daily value of stomatal resistance from day to day and the standard error bars indicate there is a lot of variation within each day (Figures 4.08 and 4.09). The maximum values are greater in 2002 but a negative surface resistance was calculated for some days.

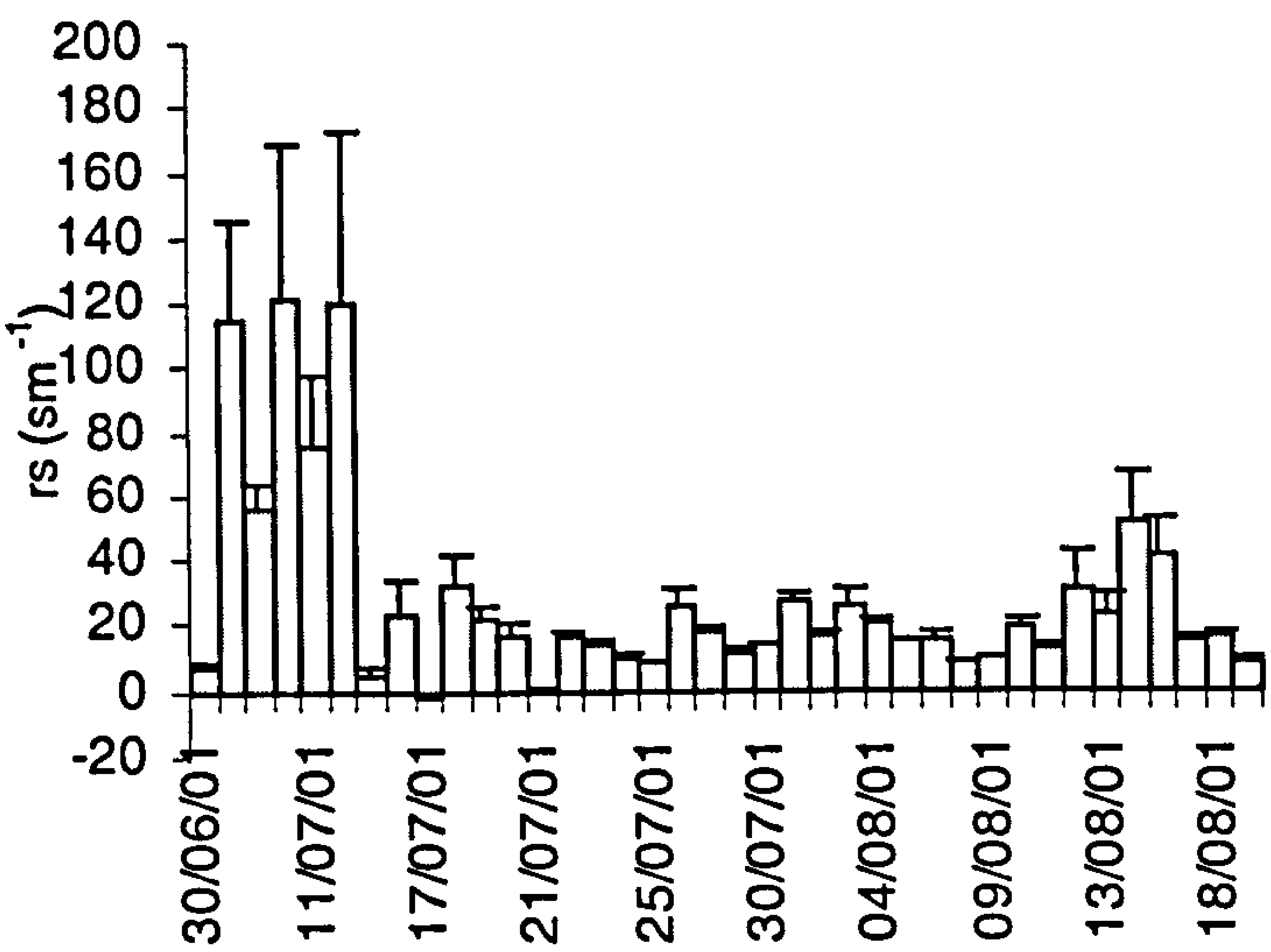


Figure 4.08 – Mean daily surface resistance, determined using the top-down approach in 2001

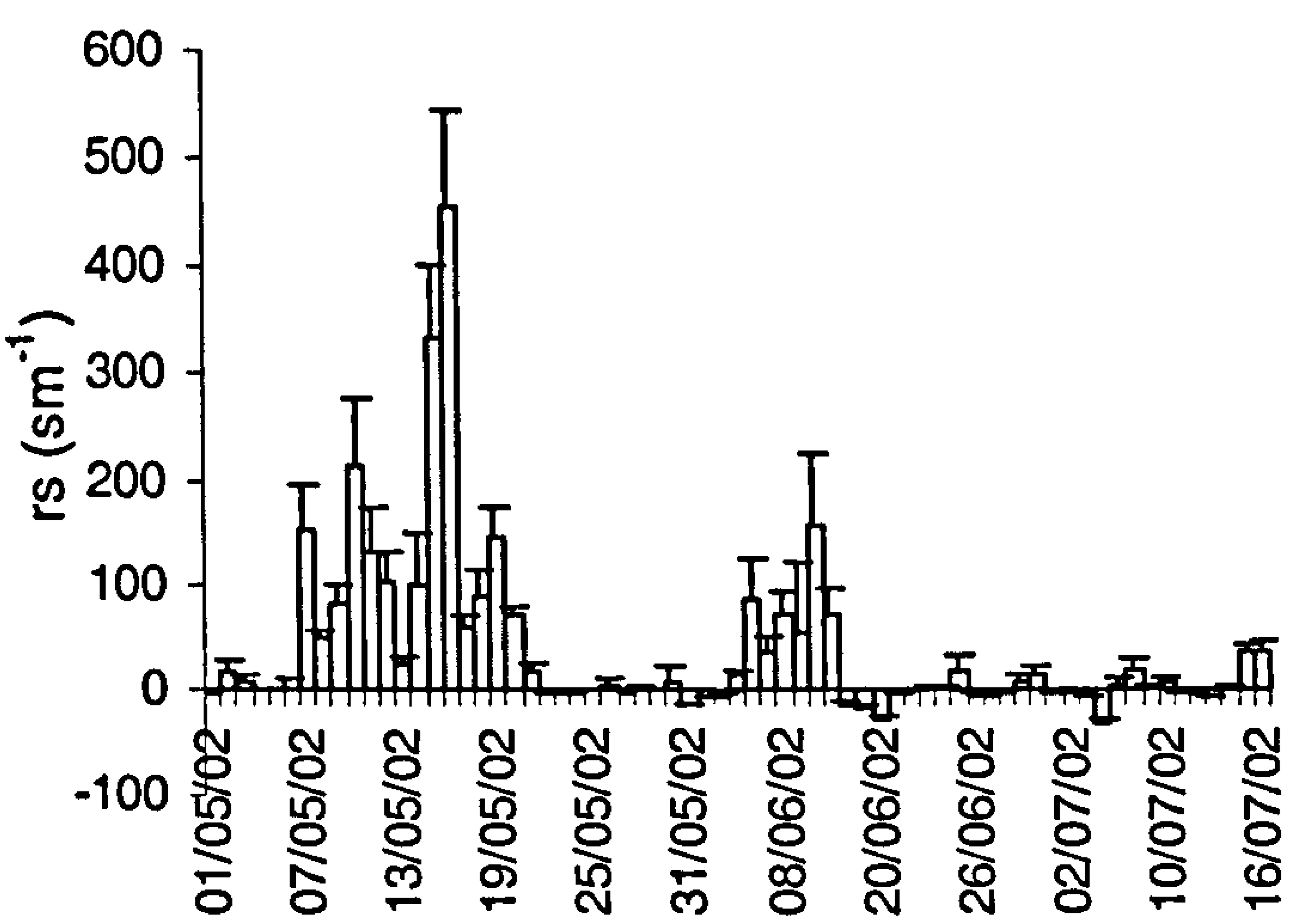


Figure 4.09 – Mean daily surface resistance, determined using the top-down approach in 2002

4.4.3 Determination of surface resistance – bottom up approach

4.4.3.1 LAI measurements

The measurement of the angle of leaves on the stems of *Phragmites* resulted in 43 leaves assessed as being nearer horizontal and 61 leaves being nearer vertical.

$$ELADP = \frac{\pi(43)}{2\times 61} = 1.10 \tag{4.15}$$

The 2001 LAI results are shown in Table 4.02.

Table 4.02 – LAI results

	20/06/01	18/07/01	15/08/01
Mean	3.39	5.51	5.71
SD	1.13	0.88	1.55
n	20	24	20

4.4.3.2 Measuring stomatal conductance

Table 4.03 summarises the mean daily results for each day measured using the IRGA and also gives the mean weather parameters for the measurement period during the day. Measurements are given in terms of a stomatal conductance, which is the inverse of resistance. Between 160 and 240 measurements were taken each day. On 29/8/01 measurements were only taken between 10:40 and 14:40 and the weather was sunny. Measurements were taken 09:00 to 17:00 on the remaining days. 28/05/02 was sunny for the first hour of measurements then remained cloudy for the rest of the day. 03/06/02 also was sunniest in the morning and became cloudier. 11/07/02 was fairly sunny all day with passing clouds apart from rain around 13:00, which led to the loss of data between 13:00 and 14:00 due to the wet canopy.

Table 4.03 – Summary of stomatal conductance results and mean weather parameters over the periods during which stomatal conductance was measured (on 29/08/01 measurements were only taken between 10:40 and 14:40) (VPD is vapour pressure deficit)

	R_n ($W\ m^{-2}$)	T (°C)	VPD (kPa)	u ($m\ s^{-1}$)	Stomatal conductance ($m\ mol\ m^{-2}\ s^{-1}$)			
	Mean	Mean	Mean	Mean	Mean	Min	Max	S.D.
29/08/01	450.7	20.4	0.86	2.6	171	98	254	29.2
28/05/02	311.5	14.4	0.38	4.9	99	27	223	37.6
03/06/02	286.5	19.1	0.46	2.2	91	20	250	40.0
11/07/02	431.6	18.8	0.40	3.2	119	33	244	40.1

The highest values were measured in August 2001 when net radiation was highest, and the lowest were measured at the end of May and the start of June when net radiation was lowest. The highest stomatal conductance occurred at the top of the canopy (Figure 4.10). However the leaves from lower down also played a significant part in total canopy transpiration

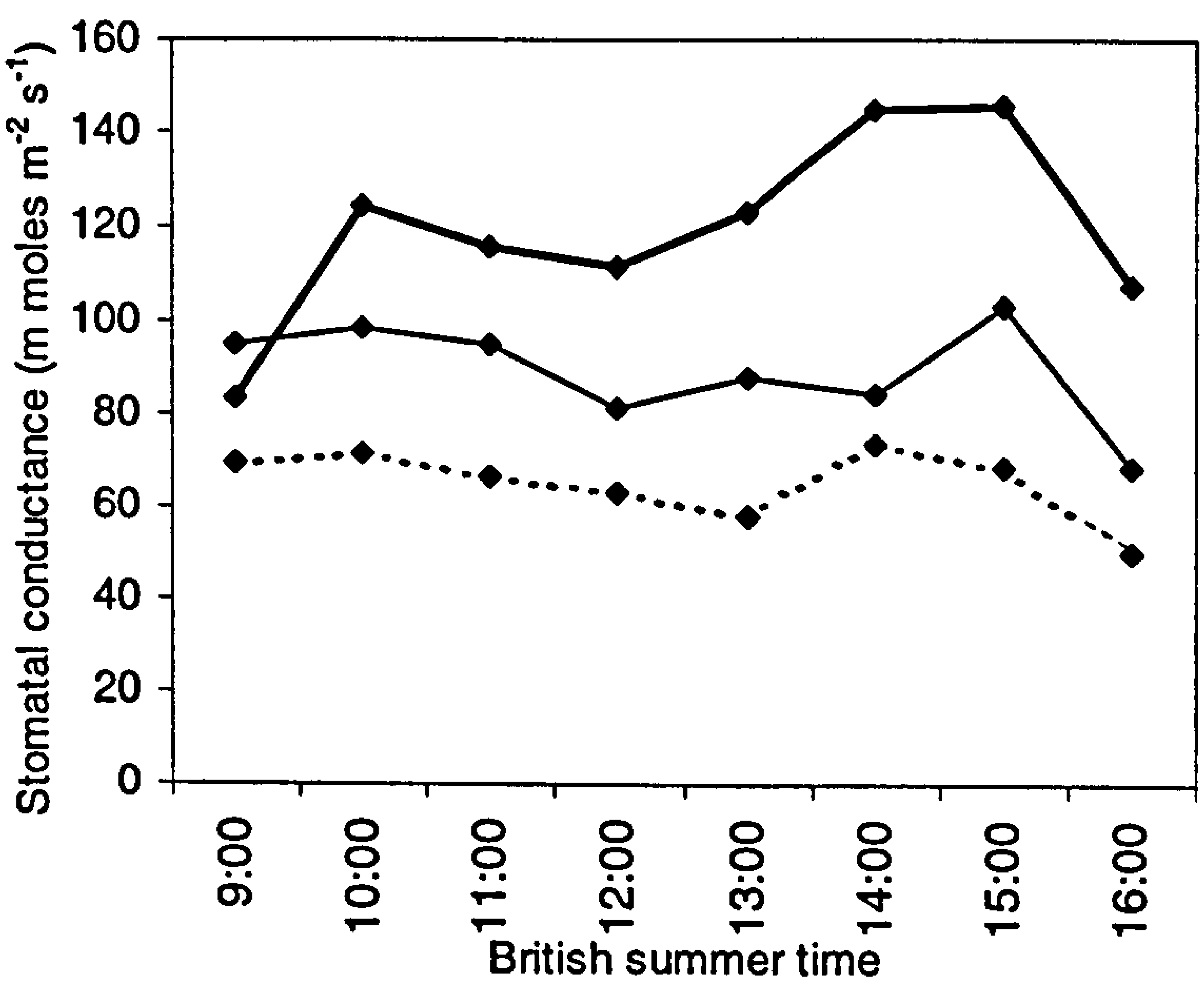


Figure 4.10 – Mean hourly stomatal conductance data for 03/06/02 showing separately the stomatal conductance measured at the top of the canopy (bold solid line), in the middle of the canopy (light solid line) and at the bottom of the canopy (dashed line).

The stomatal conductance of reeds is fairly constant throughout this cloudy day on an hourly basis. Stomatal conductance also was affected by the sampling group that it came from. A different set of leaves was used every 20 minutes and the mean of these three data sets was used to create the hourly values above. However if the three data sets are analysed separately it can be seen that at least until mid afternoon one sampling site has consistently higher values of stomatal conductance than the other two (Figure 4.11). Although sampling was done in the middle of the reedbed, inevitably a small clearing must be made in which to set up equipment and stand to take measurements. The sampling set represented by the open circles was on the south side of the clearing and therefore the dew dried first on this side (hence the earlier sampling) and it appears to have higher conductance throughout the morning and early afternoon.

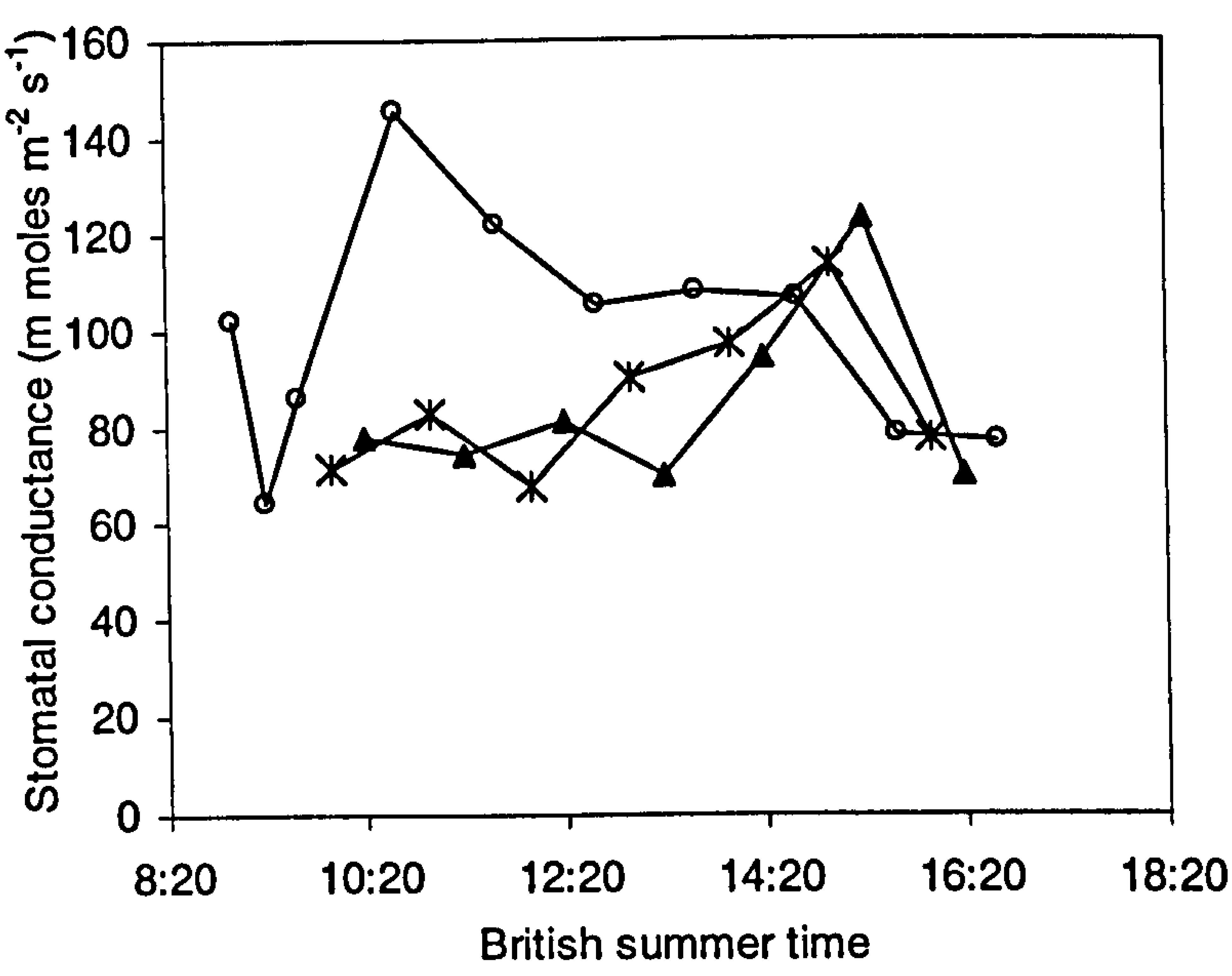


Figure 4.11 - The mean of all three levels for each of the three sampling sites used in consecutive twenty minute periods each hour. Each sampling site is shown with a separate symbol.

Figure 4.11 is for the mean of all three levels. When the three levels were analysed separately it was found that there was a pronounced difference in stomatal conductance measurements between the sampling sites for the lower and middle layers but little difference in the upper layer. This could be because the sampling procedure is allowing extra light to penetrate to the inside of the canopy on the south side but the upper layers are already exposed to sunlight so little difference is seen.

4.4.3.3 Relationships between evapotranspiration and stomatal conductance

In order to investigate whether stomatal conductance of reeds is an important factor in determining the evapotranspiration of the reedbed, the relationship between stomatal conductance as measured using the IRGA and evapotranspiration as estimated using the Bowen ratio Energy Balance Method (BREB) was investigated. For each day sampled, the twenty minute stomatal conductance values (mean of all levels) were correlated with the corresponding 20 minute evapotranspiration measurements from BREB and the r^2 values calculated (Table 4.04). The correlations were also performed with the data from each level and the ET data. The table also shows the canopy level with the strongest correlation.

Table 4.04 – The r^2 values of the correlations between measured stomatal conductance and ET measured using BREB. The table shows the result for the mean of all the measurements over the twenty minute averaging period and the result for the level that had the strongest relationship of the three levels measured. Statistically significant relationships are highlighted in bold.

	r^2 for mean r_s of all levels	Level with strongest r^2	
29/08/01	**0.387**	n/a	
28/05/02	**0.434**	middle	**0.471**
03/06/02	0.122	high	**0.529**
11/07/02	**0.171**	low	**0.297**
All days	**0.359**	low	**0.426**

All days had statistically significant relationships between stomatal conductance and evapotranspiration.

All days had statistically significant relationships between stomatal conductance and evapotranspiration.

4.4.3.4 Scaling up from leaf to canopy level

Measured stomatal conductance was converted to stomatal resistance in order to input it into the Penman Monteith equation. The conversion from conductance in m moles m^{-2} s^{-1} to resistance in m s^{-1} was made following Jones (1992) who gives a table of conversion factors (p.357). The conversion varies with temperature according to the equation:

$$y = 0.0005T^2 - 0.1586T + 43.997 \tag{4.16}$$

The measured IRGA conductance was divided by the appropriate conversion factor (y) for the temperature to find conductance in mm s^{-1}. The inverse was then found in order to convert it to a resistance (s mm^{-1}) and the result was multiplied by 1000 to give values in the same units as those in the Penman Monteith equation (s m^{-1}).

The measured stomatal resistance for a single leaf, as measured using the IRGA, was combined with the LAI values in order to create a value for surface resistance using Equation 4.03. The similarity between top down and bottom up was improved by altering the denominator of Equation 4.03. This is simply an empirical figure designed to account for the fact that only one side of the leaf takes part in transpiration. However as *Phragmites* is amphistomatous a larger denominator in the equation is reasonable and the standard figure of 0.5 was replaced by 0.7 so the scaling up equation used was:

$$r_s = \frac{r_l}{0.7LAI} \tag{4.17}$$

4.4.3.5 Comparison of top down and bottom up stomatal resistance

Bottom up and top down surface resistance were compared. On 29/08/01 stomatal resistance estimated using the bottom up approach was of a similar magnitude to that estimated from the top down approach, with the exception of a couple of data points where the top down approach estimated a higher resistance (Figure 4.12).

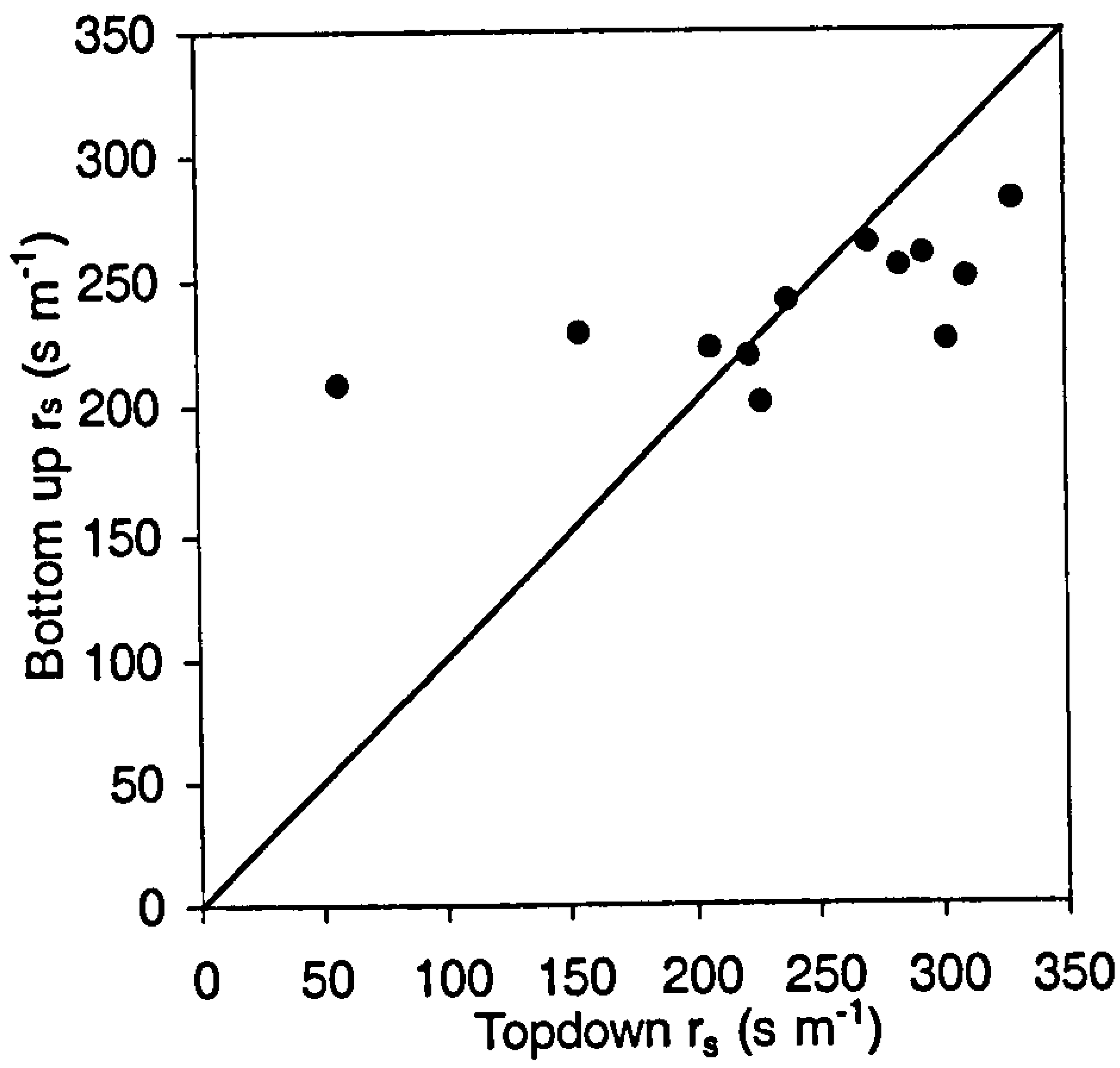


Figure 4.12 – Comparison between measured and "top down" stomatal resistance on 29/08/01

However the similarity between bottom up and top down stomatal resistance was not so good on the other days that were measured. The fit on 03/06/02 is very poor and the top down resistance is consistently much smaller than that measured, to the point that in a number of cases top down stomatal resistance values are negative. The top down values are much more variable throughout the day than those that were measured using the IRGA in the bottom up approach (Figure 4.13).

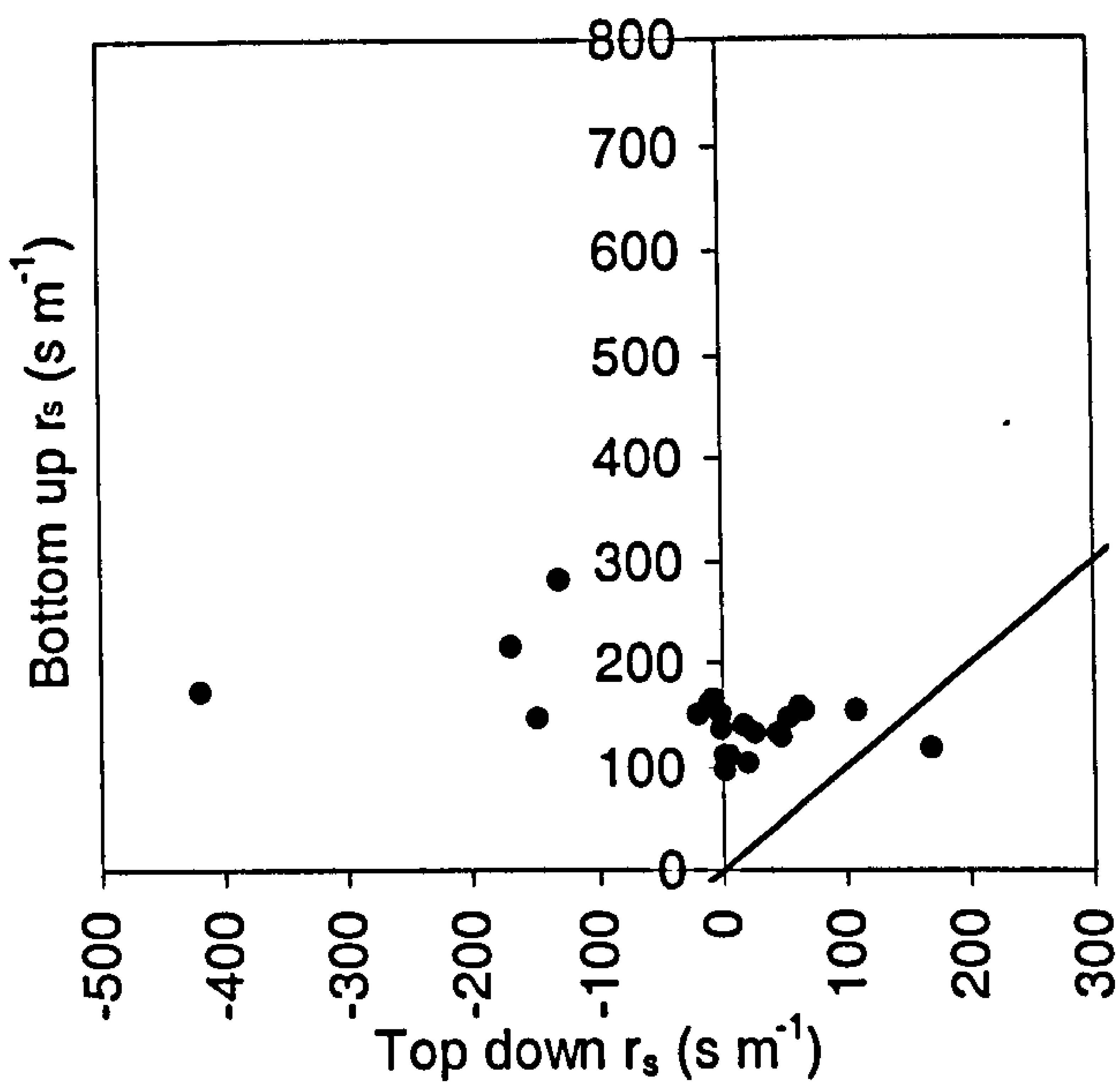


Figure 4.13 – A comparison between measured and "top down" stomatal resistance on 03/06/02

As a general rule the top down conductance values were higher than those that were measured. There has been much discussion in the literature about the relationship between top down r_s and bottom up measured r_s. Results similar to both Figure 4.12, where good relationships are found and Figure 4.13, where relationships are poor are reported. The discussion centres around the fact that bottom up conductance is concerned purely with transpiration, whilst the top down approach includes other factors. In this study the aim is to quantify total evapotranspiration. In a wetland environment there may be a significant difference between transpiration and evapotranspiration rates as the waterlogged soils and areas of open water will contribute significantly to overall evapotranspiration. The IRGA also gives an estimation of transpiration from the individual leaf which can be compared with evapotranspiration from BREB in order to illustrate this difference. In three out of four days there was a good correlation between transpiration estimated by the IRGA and evapotranspiration estimated by BREB (28/05/02 – r^2 = 0.52, 03/06/02 = 0.67, 11/07/02 = 0.47, 28/08/02 = 0.21). It can be seen that on 03/06/02 although there is a good relationship the evapotranspiration is around twice as high as the transpiration (Figure 4.14), indicating that evaporation does appear to be occurring from additional sources other than the leaves. However although on 28/08/02 the correlation is not as good, the actual magnitude of transpiration and evapotranspiration are quite similar indicating a high proportion of water is coming from transpiration (Figure 4.15).

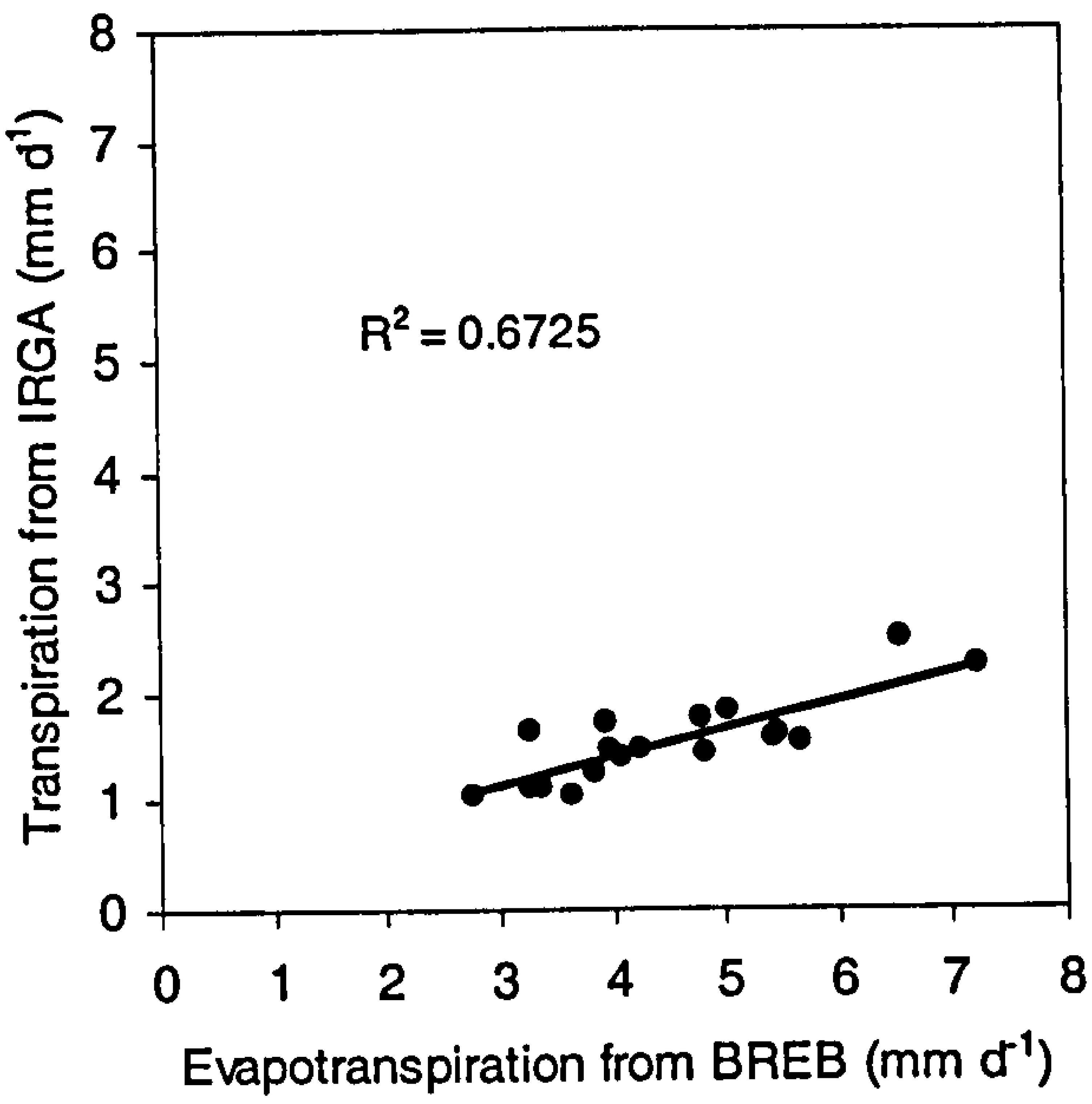


Figure 4.14 – A comparison of transpiration as calculated by the IRGA and evapotranspiration from BREB on 03/06/02.

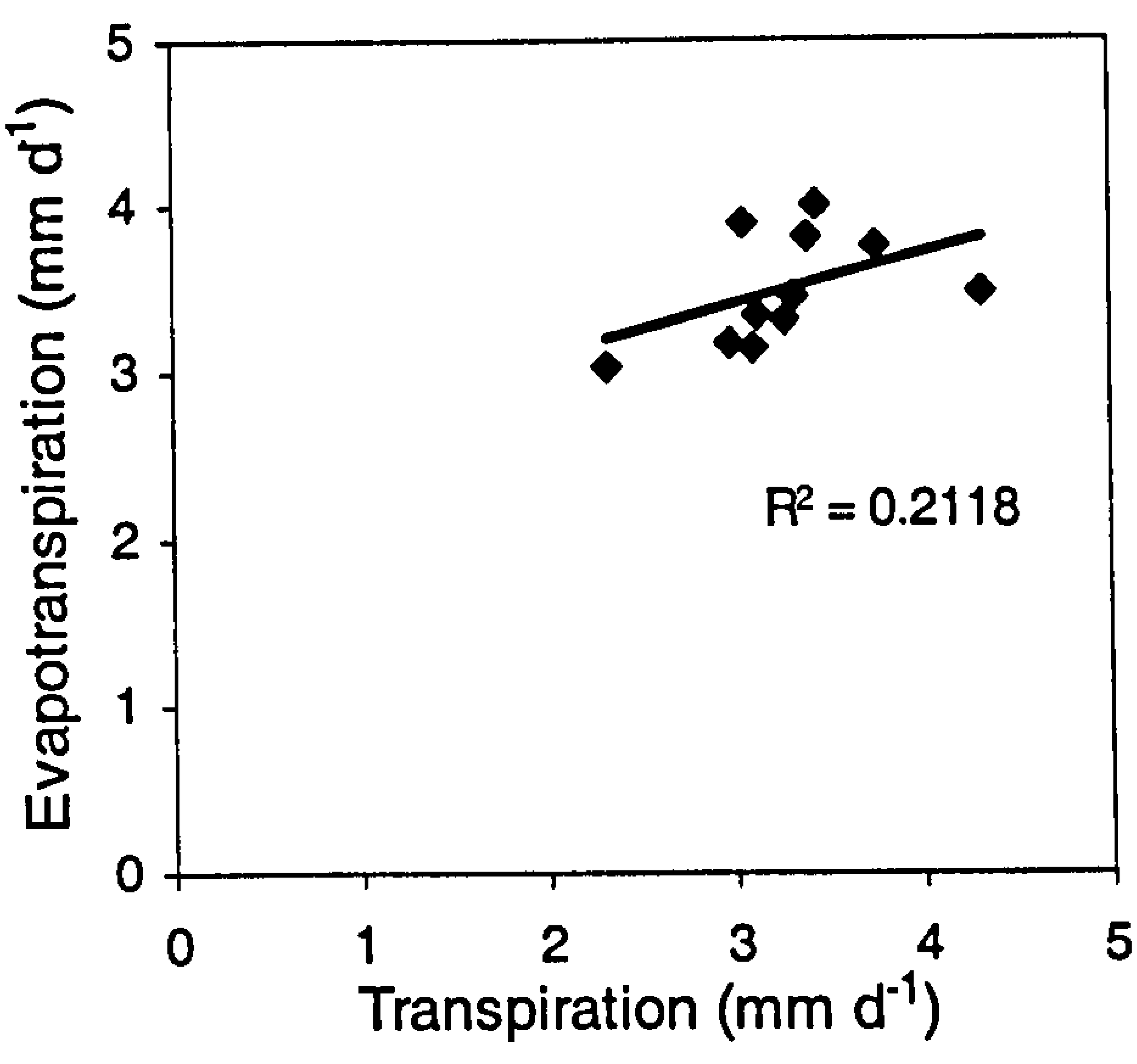


Figure 4.15 – A comparison of transpiration as calculated by the IRGA and evapotranspiration from BREB on 28/08/02

4.4.4 Modelling stomatal resistance

The aim of this section is to create a model of stomatal resistance that simply uses data from a meteorological station, therefore CO_2 levels and leaf water status are not included, although they are commonly used as inputs to stomatal resistance models.

4.4.4.1 Two step stomatal resistance model

To clarify the effect of surface resistance on evapotranspiration estimations created by the Penman Monteith equations, twenty minute Penman Monteith data using the mean calculated r_a for reed (d = 1.005 m, z_o = 0.48 m) was plotted out over the course of a single day against BREB estimated ET_{reed} for the same period. A value of r_s of 70 s m^{-1} was used following the reference equation of Allen *et al.* (1994c). If it is assumed that the mean measured values of r_a are correct, then deviation from the 1:1 line between Bowen ratio evapotranspiration and that modelled by Penman Monteith should be caused by deviations in r_s. Figure 4.16 show Penman Monteith data plotted against Bowen ratio data for three separate days.

On dry days (Figure 4.16 a, b) the data falls on a characteristic curved best fit line which falls above the 1:1 line, and in some cases crosses it at the higher evapotranspiration rates. In the majority of cases the Penman Monteith estimation exceeds the Bowen ratio estimation indicating that stomatal resistance is overestimated but when evapotranspiration is highest the Bowen ratio evapotranspiration data is greater than that from Penman Monteith. However on wet days the Bowen ratio data is much closer to that from Penman Monteith, with some data falling above and some below the line. These patterns held for most days studied.

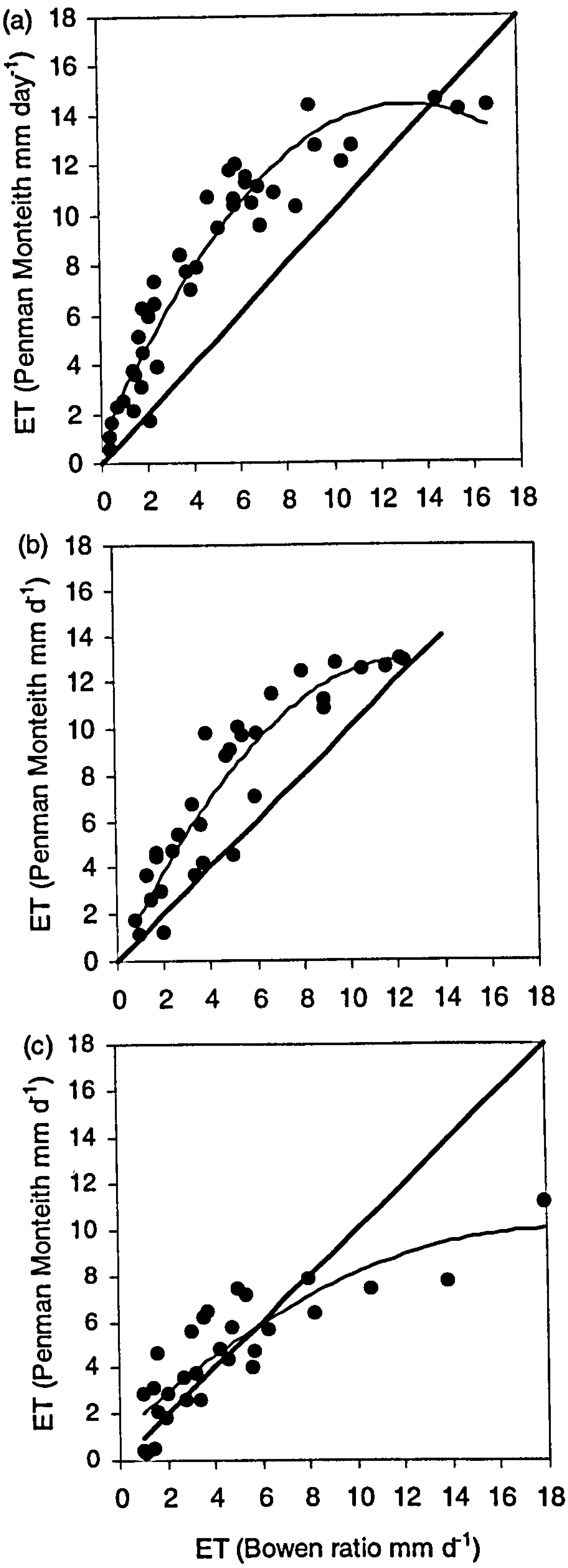


Figure 4.16 – Evapotranspiration calculated from the Penman Monteith equation plotted against evapotranspiration calculated from BREB. (a) 26/06/01 – no rainfall, (b) 23/06/01 – no rainfall, (c) 15/06/01 – 4.5 mm rainfall.

In order to begin to model this effect, stomatal resistance was altered. Initially for dry days two values were used – a higher one for when evapotranspiration was low and a lower one for when evapotranspiration was high. This was not done because it was thought to be a representation of the actual resistances experienced by the plant, but because it appeared that it may result in a better model of evapotranspiration than using a single value, whilst keeping the model simple. The definitions of "high" and "low" were assigned by eye, by observing the rate of evapotranspiration that marked the division between under and over estimation of evapotranspiration. The two values were chosen to give the best possible fit for that day. The results of doing this for the two dry days seen above are shown in the graphs below. The curves are the best-fit lines with the original stomatal resistance values and the straight lines are the new best-fit lines.

On the dry days this results in an improvement in the fit around the 1:1 line (Figure 4.17 a, b). As discussed in Chapter 3 on wet days when there is water on the leaf surface the stomatal resistance is effectively zero so the r_s value in the Penman Monteith equation applying to wet days was set to zero (Figure 4.17c). This also results in an improvement but it is not as dramatic.

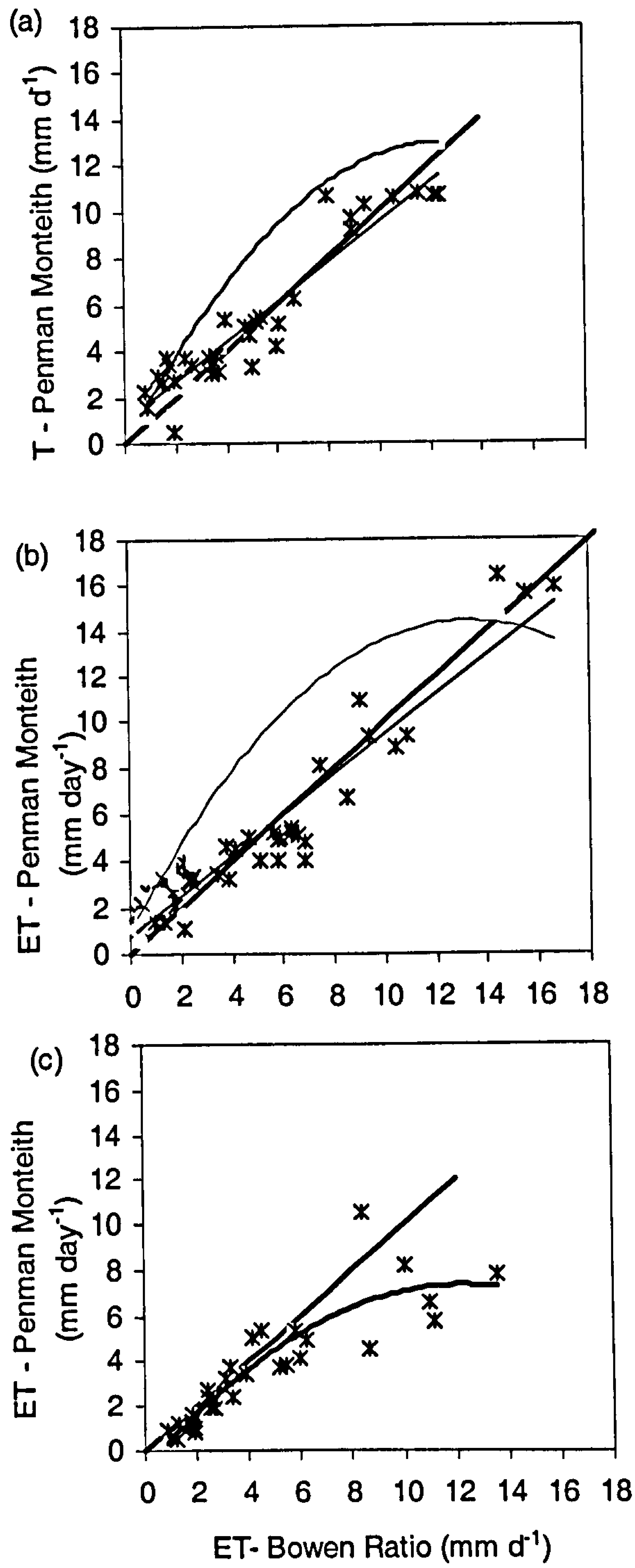


Figure 4.17 – A plot of 20 minute evapotranspiration as calculated using the Bowen ratio approach against evapotranspiration using the Penman Monteith equation. Stomatal resistances are altered so that two values are used per day. (a) 26/06/01 – a dry day using 60 and 180 s m^{-1} as resistances, (b) 23/06/01 – using 112 and 56 s m^{-1} as resistances and (c) 06/06/01 – a wet day with r_s set to zero (straight line is 1:1 line, curved line is original best fit line)

Using this technique of a two stage stomatal conductance can result in a very good match between the evapotranspiration measured using the Bowen ratio approach and that modelled using Penman Monteith averaged on a daily basis (Figure 4.18).

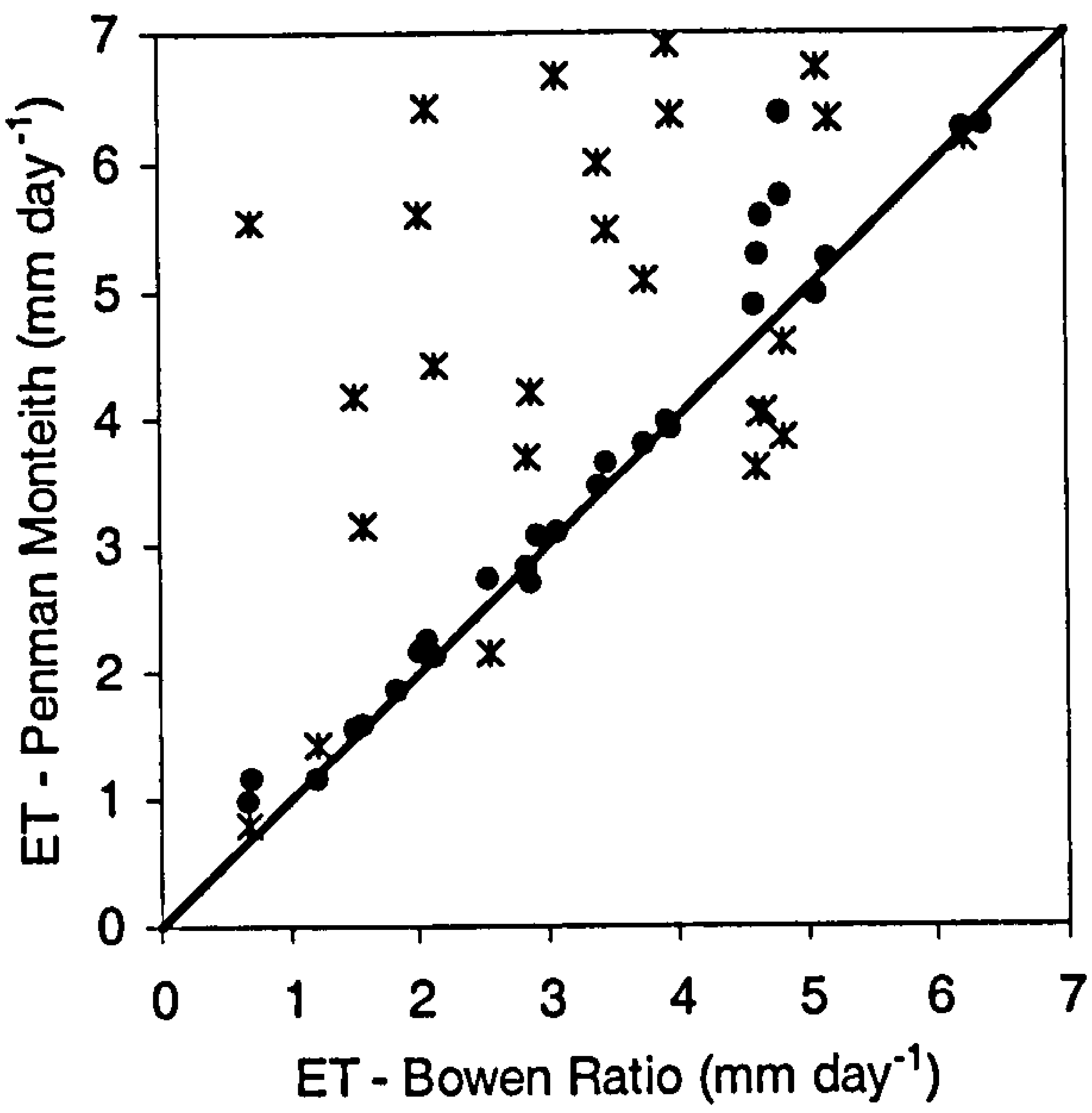


Figure 4.18 – Daily averages of data with manually altered "two step" r_s values (filled circles), and the unaltered data with r_s of 70 s m^{-1} (crosses).

There are still some days that do not fit. These are all wet days where the resistance was set to zero throughout the day. Although this brought improvement from the original there are obviously still some problems. As we are principally interested in daily average values for modelling the above graph looks promising. It shows that at least in principle it is possible to model the evapotranspiration as calculated by the Bowen ratio approach using the Penman Monteith equation. However the stomatal resistances here were fitted to the Bowen ratio data and the values necessary to get a fit on each day were not the same, presumably due to different prevailing meteorological conditions. In order to use this approach it is necessary to model the stomatal resistances from the meteorological data which causes the change in stomatal resistance to occur.

4.4.4.2 Modelling from bottom up stomatal conductance data using regression with meteorological parameters

The relationships between measured stomatal conductance and environmental variables thought to influence stomatal conductance (net radiation, temperature, wind and vapour pressure deficit) were investigated. If strong relationships exist, these could be a simple way to simulate the variation in stomatal resistance within and between days. Correlations were performed both on the average of each canopy level measured and on the total average for each twenty minute period and the square of the Pearson product moment correlation coefficient (r^2) was found. Table 4.05 shows r^2 values of correlations between measured stomatal conductance and meteorological variables. Those that are statistically significant at the 0.05 level are shown in bold typeface.

Table 4.05 –The r^2 values for the correlations between measured stomatal conductance and meteorological variables on each day. The table shows the result for the mean of all the measurements over the twenty minute averaging period and the result for the level that had the strongest relationship of the three. Statistically significant relationships are highlighted in bold.

		29/08/01	28/05/02	03/06/02	11/07/02
R_n	r^2 over all 3 levels	0.103	**0.244**	0.097	**0.226**
	Level with highest r^2	-	low	high	low
	r^2 of this level	-	**0.456**	0.097	**0.237**
Temp.	r^2 over all 3 levels	0.034	**0.326**	0.118	0.056
	Level with highest r^2	-	low	high	low
	r^2 of this level	-	**0.439**	**0.338**	0.091
Vapour pressure deficit	r^2 over all 3 levels	0.033	0.003	0.042	0.032
	Level with highest r^2	-	high	high	low
	r^2 of this level	-	0.005	**0.337**	0.147
Wind	r^2 over all 3 levels	0.000	0.033	0.031	**0.203**
	Level with highest r^2	-	low	high	high
	r^2 of this level	-	0.112	**0.172**	**0.279**

In general the correlations are not strong. It does not appear that there are simple linear relationships between stomatal conductance and meteorological variables. It would be expected that the leaves near the top of the canopy would respond most strongly to meteorological variables but this is not shown by the results. The data from all four days were combined and overall regressions found (Table 4.06). There are significant (though still fairly weak) correlations with net radiation and vapour pressure deficit. The best correlations are with the leaves measured furthest down in the canopy.

Table 4.06 – The r^2 of the combined measured stomatal conductance data for all four days and meteorological variables.

	Low	Middle	High	Overall
R_n	**0.387**	**0.191**	**0.186**	**0.341**
Temperature	0.033	0.000	0.020	0.016
VPD	**0.320**	**0.273**	**0.144**	**0.356**
Wind	0.002	0.008	0.002	0.026

However no one factor alone has a strong enough relationship to stomatal resistance to be used alone in modelling – a more complex model is required.

4.4.4.3 Modelling bottom up stomatal resistance using the model of Herbst

The measured stomatal conductance was modelled using a simple model suggested by Herbst (1995) as it combines the two parameters with which statistically significant relationships were found:

$$g_c = \frac{aR_n}{1+bD} \qquad (4.18)$$

where D is vapour pressure deficit and a and b are empirical parameters. Least squares analysis resulted in the parameters being set as a = 0.23 and b = –0.43. Figure 4.19 shows the result of this model compared with measured stomatal conductance.

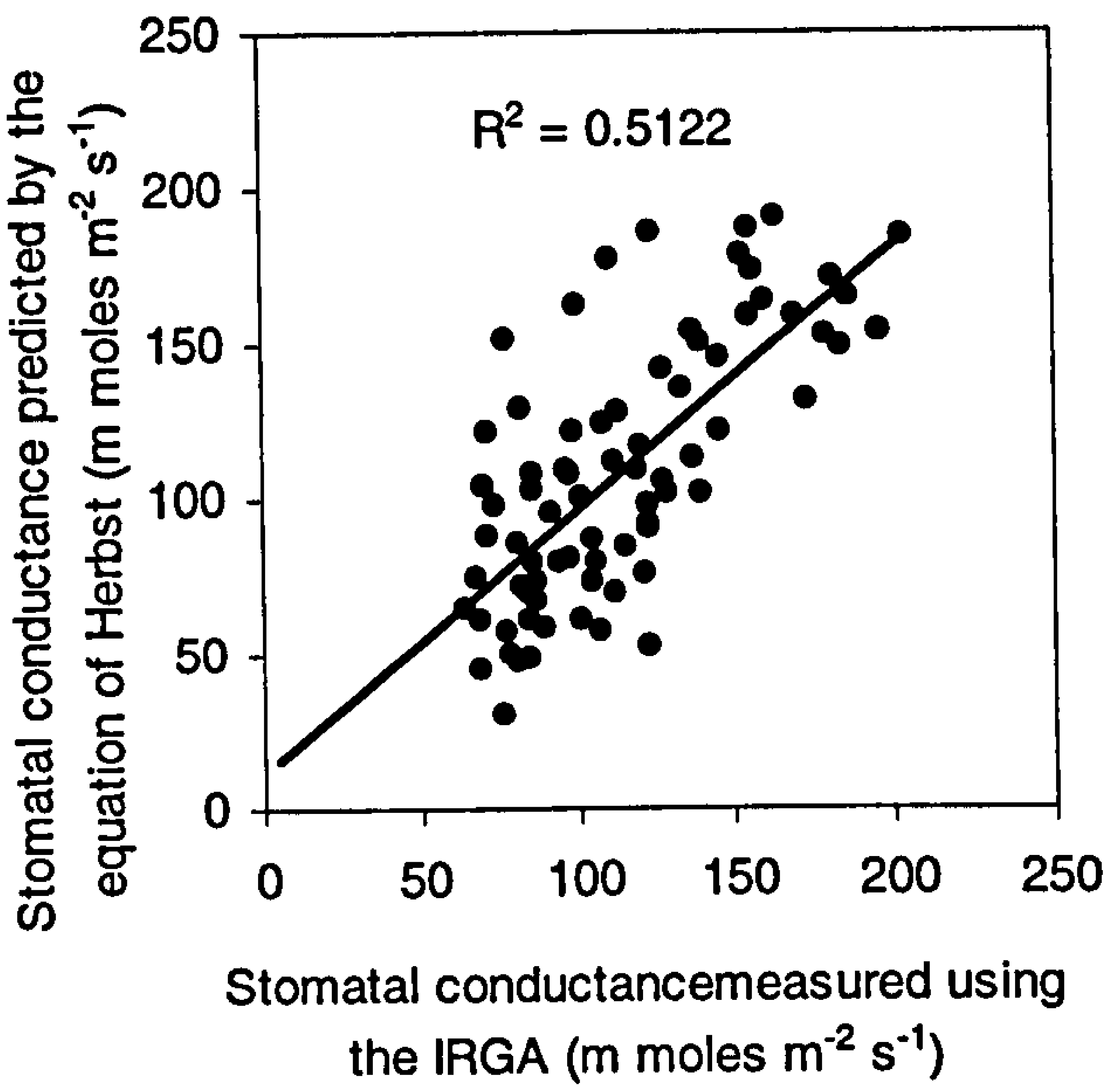


Figure 4.19 – Comparison of measured stomatal conductance with predicted values from Equation 4.18.

4.4.4.4 Modelling top down stomatal resistance using multiple linear regression

The top down data is more variable and more problematic to model than the measured data. It was not possible to fit a simple model such as the one above. A commonly used approach to modelling stomatal conductance is to use a linear multiple regression equation such as :

$$g_s = a + (b\,R_n) + (c\,T) + (d\,r_a) + (e\,D) \tag{4.19}$$

This was tried and the best fit result came out with a r^2 of 0.15 – a very poor result. A scatter plot of the top down versus predicted data can be seen in Figure 4.20.

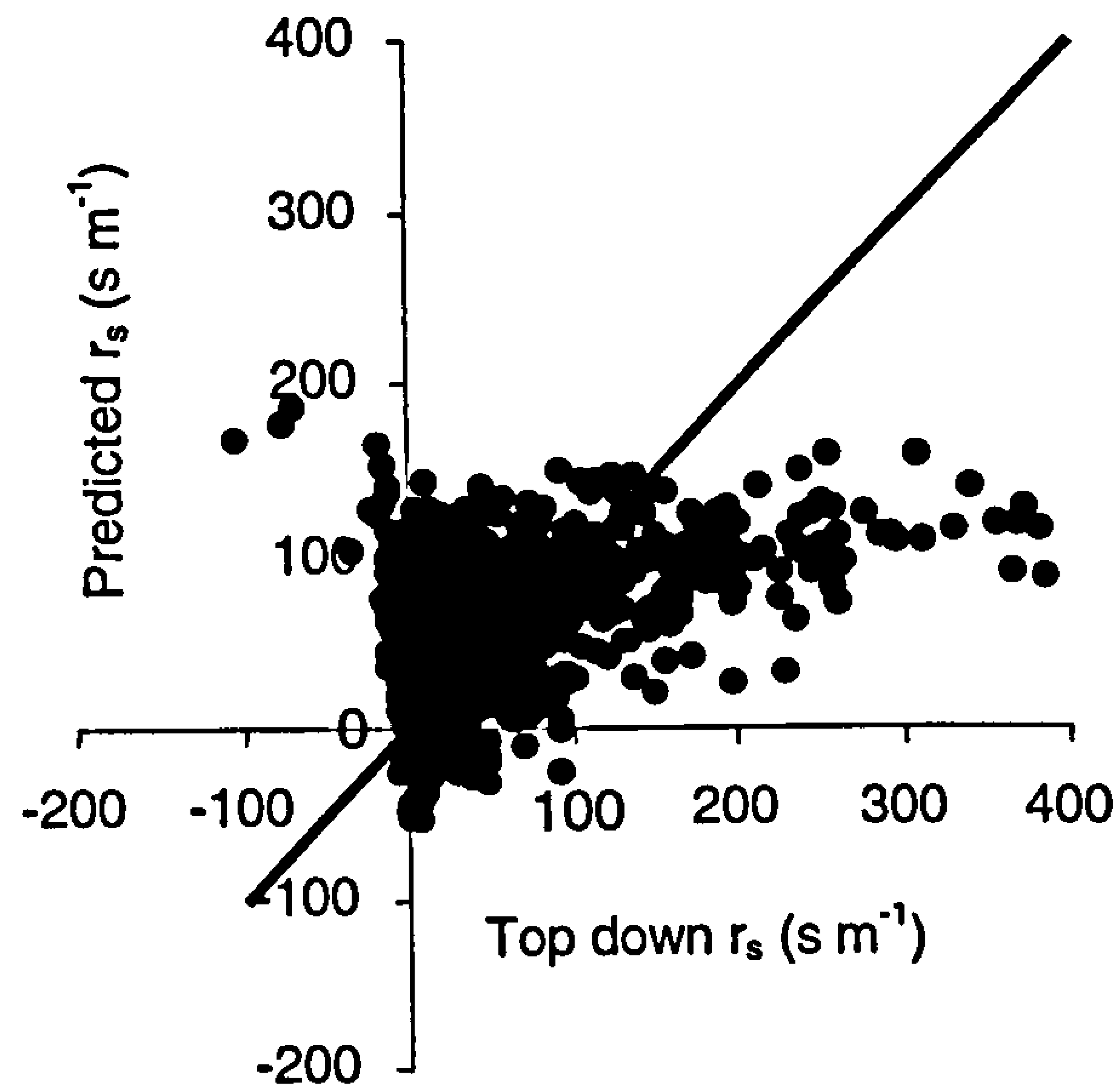


Figure 4.20 – A comparison of top down stomatal resistance with values predicted using a linear regression model

However this result is not altogether surprising since the relationship between stomatal conductance and meteorological variables is known to be far more complex than the simple linear relationship above.

4.4.4.5 Modelling top down stomatal resistance using a Jarvis-type Model

The most commonly used and recommended stomatal resistance model in the literature is that first proposed by Jarvis (1976). Stomatal resistance is expressed by a minimum resistance multiplied by a series of functions each representing one variable. The general Jarvis model is given in Equation 4.04.

The details of the determination of the functions and subsequent parameterisation are given in Appendix F. A short summary of the approach and results is given below. As top down resistance data is being used, the functional relationships are essentially determined by the form of the Penman Monteith equation and the specific Jarvis model is:

$$r_s = r_{s\min} \times \left(a + \frac{b}{R_n}\right) \times (cr_a + d) \times (e + fD) \times \left(gT^2 + hT + i\right) \qquad (4.20)$$

After parameterisation the model did not produce good results, often resulting in negative values when they should be positive (Figure 4.21), although parameterisation appeared to be successful.

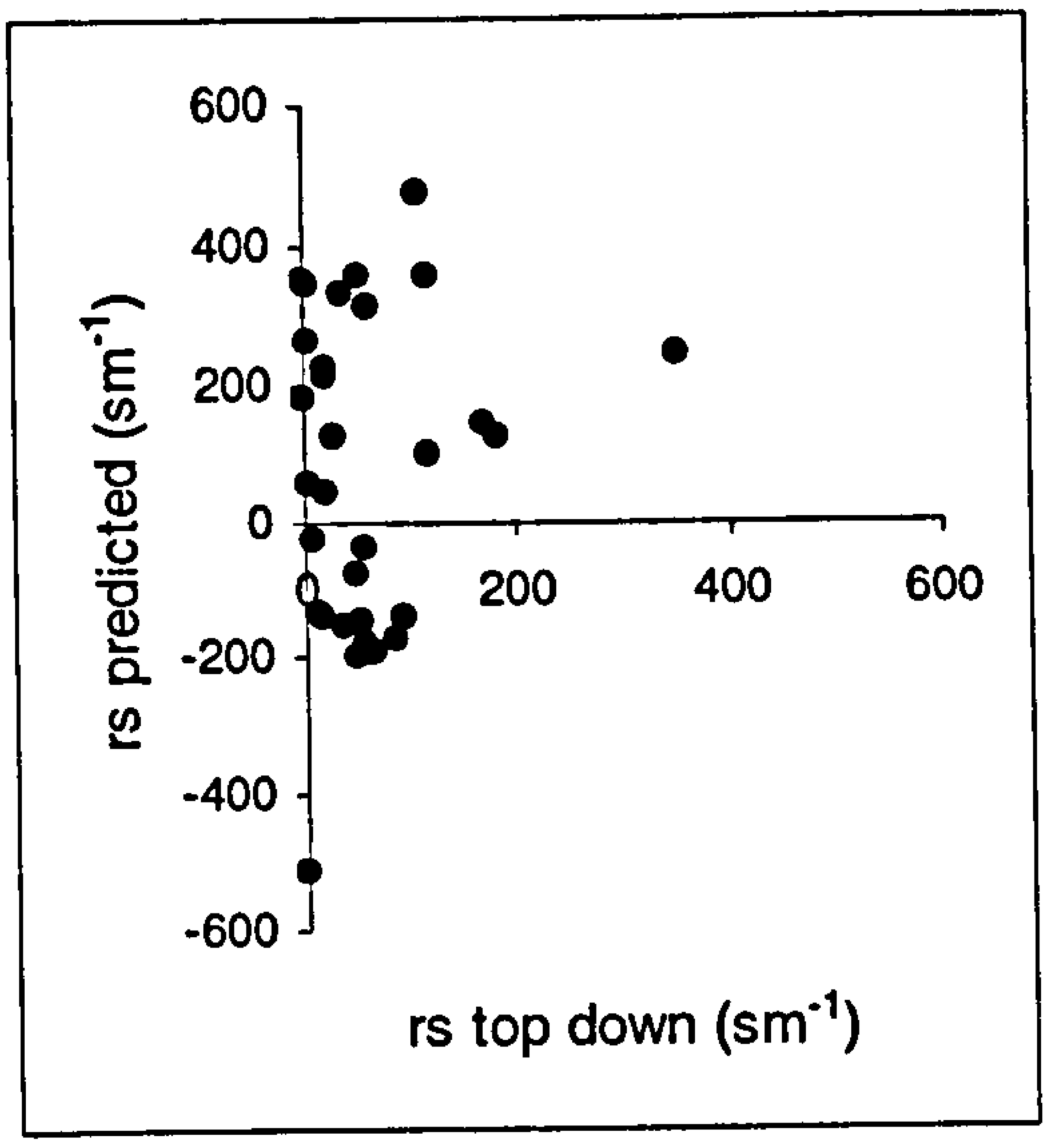


Figure 4.21 – Results of the Jarvis model prediction.

The poor result was due to the extreme sensitivity of the parameters in the function determining vapour pressure deficit.

4.4.5 One step Penman Monteith model for *Phragmites*

The parameters of the equation (d and z_o) did not change significantly throughout the measurement period and no relationship was found with vegetation height. Therefore a single seasonal average value was used for each year. These are compared with the result of using the standard reference equation in order to make clear what impact the altered aerodynamic resistance is having on the Penman Monteith estimation of evapotranspiration from reeds (Figure 4.22).

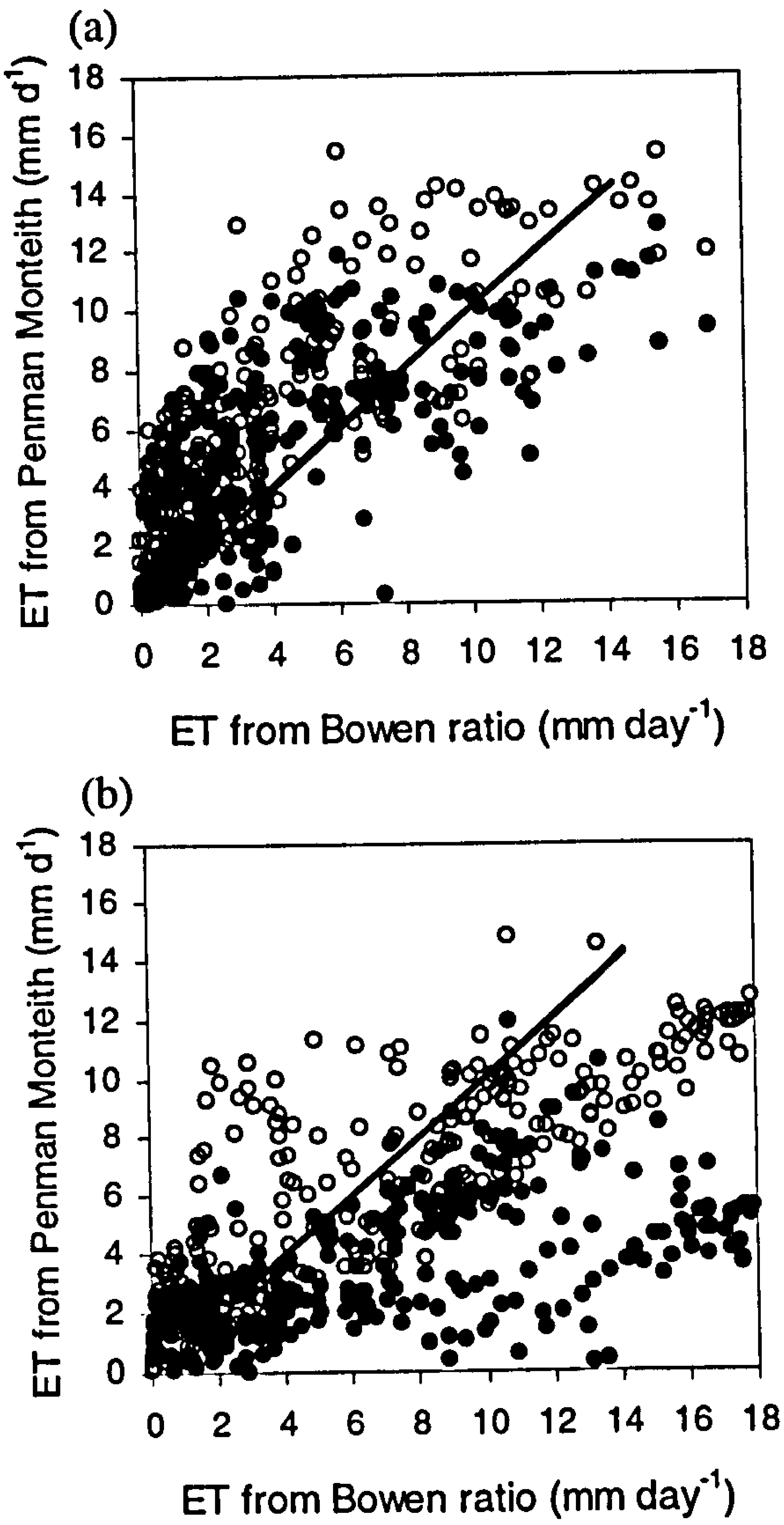


Figure 4.22 – 20 minute values using average z_o (0.48 m) and d (1.005 m) and reference values for r_s (70 s m^{-1}) for (a) 2001 (z_o = 0.48 m, d = 1.005 m) and (b) 2002 (z_o = 0.32 m and d = 1.03 m) to calculate r_a (•) compared with values using the same data set but values for r_a from the reference equation (z_o = 0.015 m d = 0.08 m) (o).

In 2001 the net result of the change in aerodynamic resistance is very little overall change in evapotranspiration magnitude between the original reference evapotranspiration parameters, and the use of the altered values calculated for reed. In 2002 there appears to be a decrease in evapotranspiration.

4.4.5.1 Including stomatal resistance in the equation

The only reasonable stomatal conductance model available is the model of Herbst (1995) using radiation and vapour pressure deficit created from the bottom up measurements. This was inserted into the Penman Monteith equation for dry days. Stomatal resistance was set at 0 for wet days (Figure 4.23).

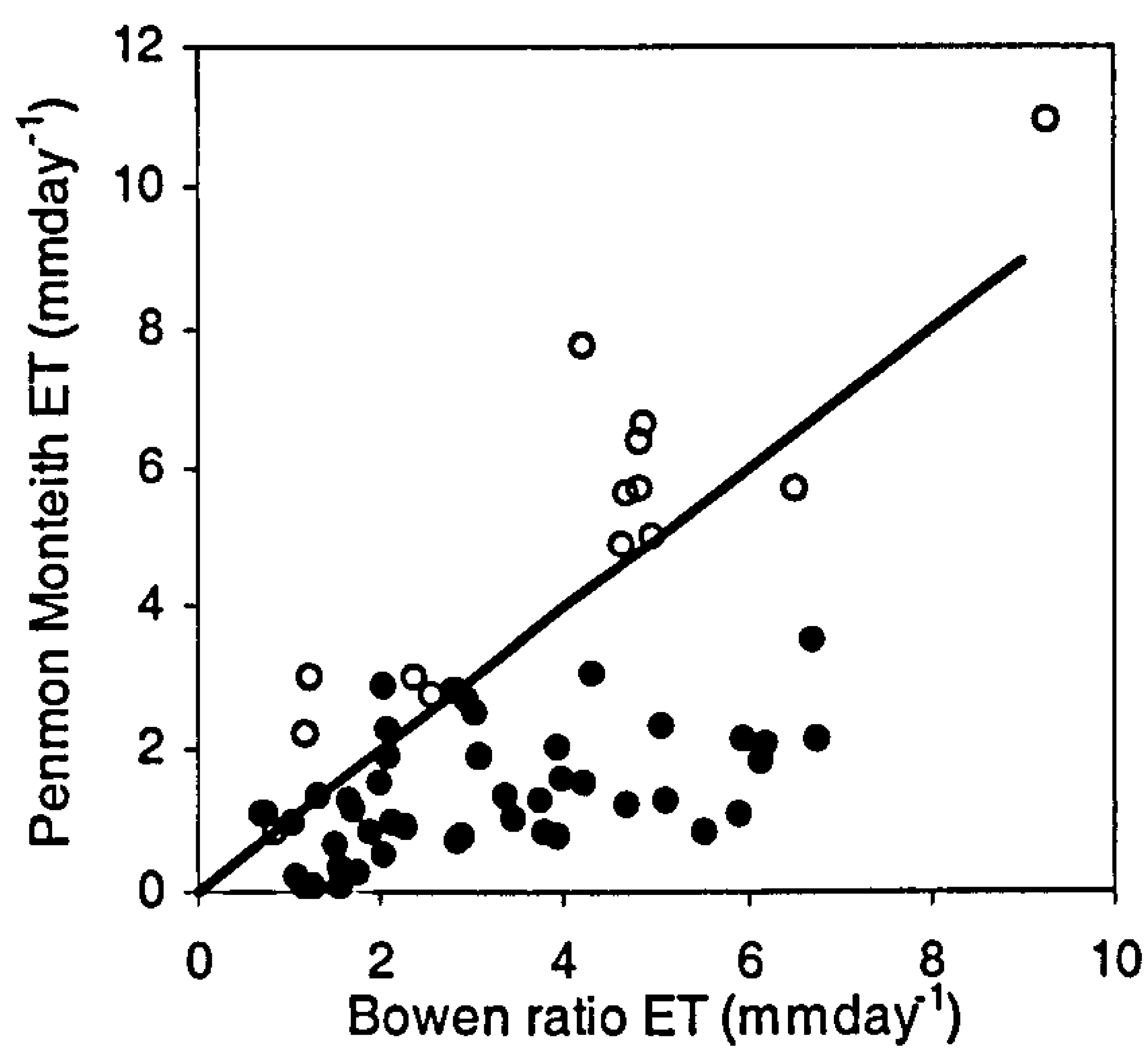


Figure 4.23 – Penman Monteith equation calibrated for reeds using average measured d and z_o values, equation 4.18 for stomatal resistance on dry days (closed circles) and 0 on wet days (open circles) for 30/05/01-28/06/01 and 22/07/01 – 28/08/01 (the intermediate period was used to calibrate aerodynamic resistance).

The model causes Penman Monteith to under-predict ET. This is unsurprising as this is a model of stomatal conductance and it has been shown that evapotranspiration from reeds is greater than the sum of stomatal conductance due to the contribution of soil and open water evaporation. The setting of wet days at zero caused over-prediction.

As the modelling of stomatal resistance has not been successful, a single figure must be used, although this is not ideal, as this is unlikely to reduce scatter. McGlinchey and Inman-Barber (1996) successfully parameterised the Penman Monteith equation for sugarcane by finding best-fit values for r_s and this approach was followed. The data were calibrated for the period 30/05/01-28/06/01 and validated for 22/07/01 – 28/08/01. A separate value was found for wet days and dry days.

The best fit values are:

Dry day: 59.4 s m^{-1}

Wet day: 23.17 s m^{-1}

The resultant calculations for evapotranspiration were compared with original reference evapotranspiration over a separate validation period (Figure 4.24).

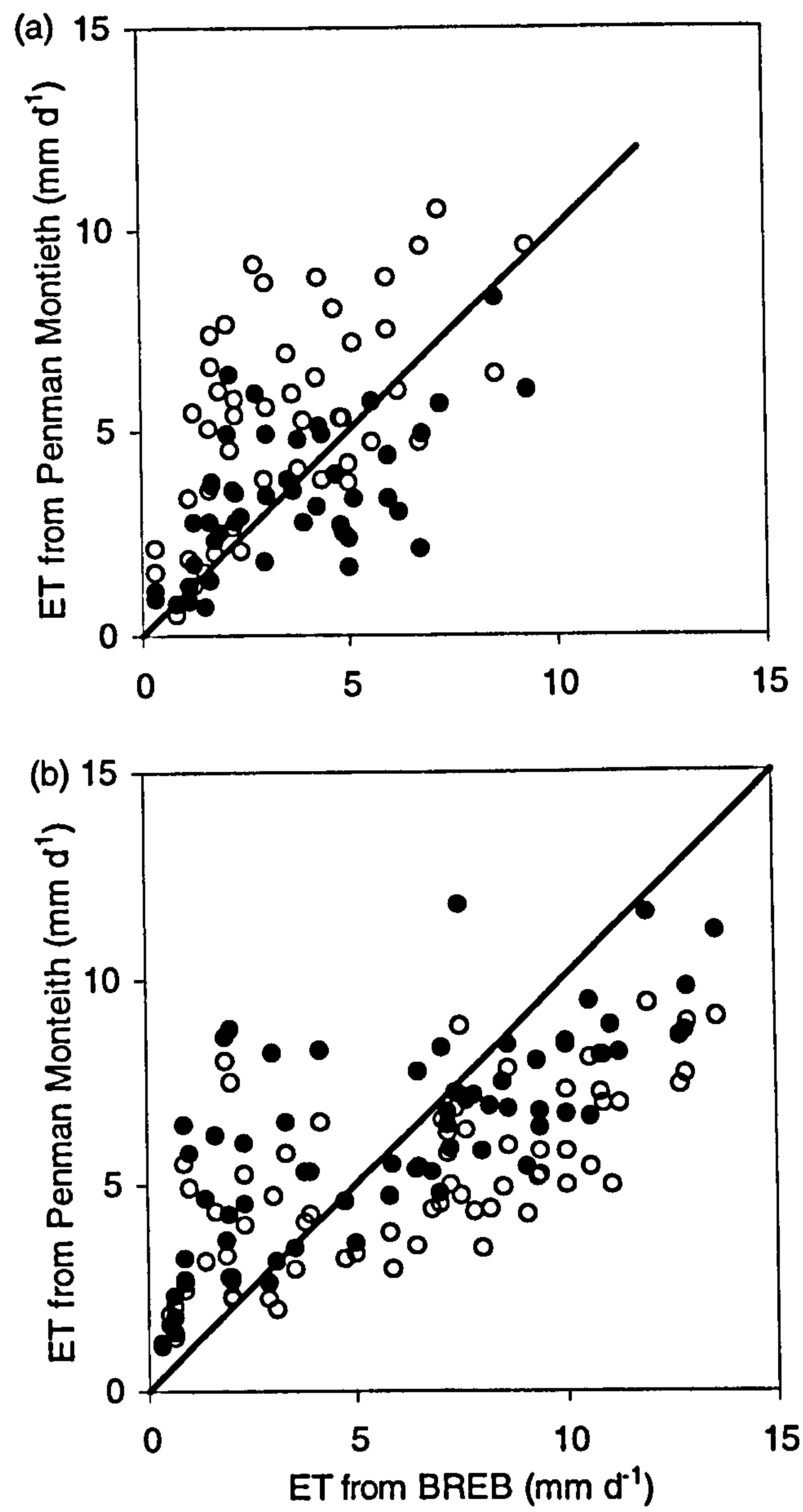


Figure 4.24 – Daily average values using average z_o and d and best fit values of r_s (•) compared with values using the same data set but values for r_a and r_s from the reference equation (o) for (a) 2001 and (b) 2002.

4.4.6 Comparison between the one-step Penman Monteith equation and the use of crop coefficients

In order to determine whether the use of the above equation results in a more accurate determination of evapotranspiration as measured using the Bowen ratio approach than would the use of crop coefficients, both approaches were used on a set of daily data from 2001 and 2002 and compared using goodness of fit statistics (Table 4.07).

The statistics used were:

Coefficient of Determination (D_c):

$$D_c = \frac{\sum (q_c - \overline{q}_c)^2 - \sum (q_c - q_{est})^2}{\sum (q_c - \overline{q}_c)^2} \qquad (4.21)$$

where q_c is observed value, $\overline{q}_c$ is the mean observed value and q_{est} is estimated value from regression line of q_c on q_e. D_c is always less than 1 and the closer to 1, the better the model.

Coefficient of Efficiency (E):

$$E = \frac{\sum (q_c - \overline{q}_c)^2 - \sum (q_c - q_e)^2}{\sum (q_c - \overline{q}_c)^2} \qquad (4.22)$$

where q_e is estimated value.

E is also less than 1. If the results are highly correlated but biased the value of E will be lower than that of D_c.

Sign tests are a simple method of testing whether there are systematic errors. A plus sign was allocated to each overestimate and a negative sign to each underestimate. A X^2 test was used to compare the number of plus and minus signs with the expected number.

Table 4.07 - Goodness of fit statistics for the crop coefficient and direct Penman Monteith models to the Bowen ratio data for 2001 and 2002.

	2001		2002	
	Crop coefficients	Direct Penman Monteith	Crop coefficients	Direct Penman Monteith
D_c	0.76	0.69	0.69	0.74
E	0.27	0.29	0.41	0.60
X^2	0.28	0.10	0.07	0.90

4.4.7 Atmospheric coupling results

Table 4.08 shows the results of the coupling factor for July 2001.

Table 4.08 – Results of the coupling factor July 2001

	Average Ω	S.D.
20 minute average	0.298	0.250
daily average	0.316	0.222
wet	0.423	0.240
dry	0.236	0.177

The value is roughly what would be expected as it is between the published values for heathland and cotton. The reedbed is fairly well coupled to the atmosphere (0 – perfectly coupled, 1 – decoupled). This indicates that the reedbed is a rough surface and stomatal conductance is likely to be important. We would expect a reduction in coupling when the surface is wet as when the leaves are wet the leaf vapour pressure deficit becomes zero and remains so irrespective of changes in ET rate and therefore it is decoupled from the atmosphere.

4.5 DISCUSSION

The research presented in this chapter aimed to parameterise the Penman Monteith equation for a reedbed. This also allowed investigation into some of the characteristics of evapotranspiration from reedbeds and some of the reasons for the differences in evapotranspiration between reeds and a reference grass surface that were found in Chapter 3. The main difficulty in parameterising the Penman Monteith equation is in working out the aerodynamic and surface resistance parameters.

4.5.1 Determining the aerodynamic resistance of reeds

Aerodynamic resistance was calculated using a standard wind profile approach. A limitation of this technique is that the logarithmic windspeed profile is theoretically only valid when the temperature profile in the atmosphere is close to neutrality (Jones 1992). Increasing instability leads to smaller values of r_a compared to the neutral value. In addition, according to surface layer similarity theory, these flux-profile relationships are only valid for heights well above $d + z_o$ and profile measurements should not be made close to the top of the canopy. This results in discrepancies in values compared to independent measurements (Cellier and Brunet 1992). Measurements have been found to result in overestimation of aerodynamic resistance and an underestimation of z_o (Thom *et al.* 1975; Thom and Oliver 1977). In order to use wind profile measurements to calculate z_o and d correctly, measurements should be made at 2-3 times the canopy height (Cellier and Brunet 1992). However this is often impractical in the case of tall vegetation such as reeds, as fetch often becomes limiting. Although Cellier and Brunet (1992) give an alternative wind profile equation its use is limited because it is difficult to evaluate the depth of the roughness sub-layer and the factor that describes the extent to which the profiles are distorted. In the current study the lower anemometer measurement was made at 2.5 m in a canopy that was 0.7 – 1.9 m tall. However in reeds there are always standing dead stems which are around 2 m tall in the canopy which may also distort the wind profile. It is therefore possible that there is some inaccuracy in the aerodynamic resistance measurements, although as the values of r_a are small the error should be small also. However far from finding an underestimation of z_o, the values measured were actually larger than were expected according to the height relationships presented by Brutsaert (1982):

$d/hc = 0.67$ (4.21)

and

$z_o/hc = 0.126$ (4.22)

The average vegetation height for the measurement period in 2002 was 1.49 m. The average value of d was 1.03 m and the average value of z_o was 0.32 m. Thus:

$d/hc = 1.03 / 1.49 = 0.69$

$z_o/hc = 0.32 / 1.49 = 0.21$

The overall relationship between d and vegetation height is very similar to the theoretical value whereas the relationship to z_o is quite different. The measured value of z_o is what would be expected of a crop of around 2.5 m tall. This result was also found by Harding (2000) who attributed it to the surrounding trees and buildings increasing momentum transport and hence effective roughness length. There are a few isolated willow trees in the reedbed which may be having an impact on momentum transport in addition to the influence of the previous season's standing dead reed stems.

Other authors present more complex relationships between vegetation height and d and z_o. Tanner and Pelton (1960), as cited in Shaw and Pereira (1982), found that z_o could be related to canopy height as $z_o = ah^b$ where h is vegetation height and a = 0.13 and b = 1.0. Plate and Quraishi (1965) found a to be 0.15 and Szeicz *et al.* (1969) concluded it was 0.10. Stanhill (1969) fitted $d = ah^b$ with b = 0.98 and a = 0.70. However these relationships did not fit the data used in this study. Over the growing season there seemed to be no relationship between vegetation height and d or z_o. Shaw and Pereira (1982) attribute the ineffectiveness of these formulae to the fact that they do not account for canopy density and leaf distribution. They found that increasing plant density resulted in an increase in d and found that z_o increased with density up to a peak and then declined. The reedbed is a dense canopy and this may be a factor in why the literature relationships fail to explain the results found.

The calculated values of d and z_o resulted in a value of aerodynamic resistance that is lower than that used for reference grass. Aerodynamic resistance determines the transfer of heat and water vapour from the evaporating surface into the air above the canopy. Intuitively it would be expected that a lower aerodynamic resistance would result in greater transfer and therefore higher rates of evapotranspiration. Indeed, Anderson and Idso (1987) state that smaller plants (such as grass) have greater aerodynamic resistance to evaporation and therefore a lower evaporative rate in comparison to a tall canopy. However with the Penman Monteith equation changes in aerodynamic resistance can increase or decrease evapotranspiration as r_a occurs in both the numerator and the denominator of the equation (Jones 1992). In this research it was found there were many occasions when the lower aerodynamic resistance of reeds compared with the reference surface resulted in a decreased estimation of evapotranspiration. This particularly happens when radiation is high and vapour pressure deficit is small. Thom (1975) states that when

$$\frac{1}{1+\beta} > \frac{s}{s+\gamma} \qquad (4.23)$$

λE increases with r_a. When equation 4.23 is not satisfied, the Penman Monteith equation predicts that decreasing r_a will cause λE to become smaller, as was seen in the data presented here. Therefore although it is often assumed that a tall crop with lower aerodynamic resistance will have higher evapotranspiration rates than a shorter crop, the Penman Monteith equation demonstrates that under certain conditions this may not be the case.

4.5.2 Determining surface resistance – top down

Top down daily mean values of surface resistance were very variable. The 2002 data also included a number of negative values. If r_s were exactly equivalent to stomatal resistance this should not be possible. However according to Alves *et al.* (1998) these values do have a physical meaning and indicate that the evaporating surface is above the level of $d + z_o$. This can happen if the leaves that contribute most to evapotranspiration are those at the top of the canopy, above the level of $d + z_o$.

4.5.3 Determining surface resistance – bottom up

In order to investigate the surface resistance term, the stomatal conductance of reeds was measured using an IRGA. It was found that stomatal conductance was very variable between leaves and throughout the canopy due to changing light availability, vapour pressure deficit and temperature. Not all leaves contribute equally to transpiration. Conductance was lowest at the bottom of the canopy where light levels were restricted.

The Penman Monteith equation assumes the canopy consists of a single big leaf with a single value of stomatal conductance. An appropriate total for all the leaves is therefore required. As vapour pressure deficit varies down the canopy, even those leaves with the same stomatal conductance do not necessarily contribute the same amount to overall transpiration. Models that have been created to weight different leaves according to shading, leaf vapour pressure deficit and so on are, however, complex and conductance was simply averaged within the layers. Using a simple scaling up model (Equation 4.03) the mean measured surface resistance was 98 s m^{-1}. This is greater than the value used in the reference equation, which is around 70 s m^{-1}.

Only one other study of *Phragmites* stomatal conductance was available to compare the results with. Vanyarkho (1996) measured the stomatal conductance of *Phragmites australis* in Nebraska from May to October 1994 using a Li-Cor closed loop gas exchange system and also made some measurements with a prototype open gas exchange system. At the start of the season g_s was 100-200 m mol m^2 s^{-1}. Maximum magnitudes of g_s were measured in late June to mid August with an average of 300-400 m mol m^2 s^{-1}. The rates at the end of the season were low – 50-100 m mol m^2 s^{-1} due to senescence. However during all parts of the season there was a lot of scatter. A large number of days were sampled but it appears that readings were not taken throughout the day but quoted values are only for under full sunlight. The measurements in the current research were rarely taken under full sunlight as other than the day in August the measurement days were quite cloudy. Measurements in the current study were lower at the start of the season and until early July (50-200 m mol m^2 s^{-1}). This is similar to the results of Vanyarkho (1996) for the start of their season but the maximum values measured here never reach the 300 – 400 m mol m^2 s^{-1} measured by Vanyarkho (1996), continuing to have maximum around 200 m mol m^2 s^{-1}. However our results for the end of August were slightly higher (170-250 m mol m^2 s^{-1}). This was before senescence had begun and was probably the sunniest measurement day. Differences in values however would be expected, as conductance is variable from hour to hour and from day to day. The four days sampled were not necessarily typical. (Sanchez-Carrillo *et al.* (2001) also measured stomatal conductance of *Phragmites* using a porometer but only quote transpiration values in their paper).

There is potential for error in measuring stomatal conductance. Idso and Anderson (1988) state that in many studies there has been a lack of relationship between porometer derived values of leaf stomatal conductance and actual plant water status assessed by other means. Idso and Anderson (1988) used a porometer and a portable photosynthesis system to measure the conductance of water hyacinths. They concluded that inside the chamber the stomata are induced to close because the chamber air is much drier than the usual moist air that surrounds the leaf. This drier air induces an increase in water loss so stomata close and conductance measurements may be reduced. The leaf in the chamber is then no longer representative of those in the open field. A closed system has the leaf enclosed in the chamber with a humidity sensor. Closed systems have limited applicability for transpiration measurements because increasing humidity decreases the vapour pressure gradient and therefore the transpiration (Pearcy *et al.* 1989).

Positive correlations were found between the transpiration measured by the IRGA and the evapotranspiration measured by BREB which gives us some confidence in the IRGA results. Evapotranspiration was significantly greater than transpiration and this may indicate that water vapour loss from sources other than the leaves is important. When LAI is less than two and the soil is wet, soil evaporation can have an important influence on total evapotranspiration (Saugier and Katerji 1991). The contribution of evaporation from soil may be greater when there is standing water present. The greatest difference between transpiration and evapotranspiration was found in the measurements that were taken earlier in the season when LAI is lower. They were not found in the measurements taken in late August when LAI is greater (LAI 5.5). This may be linked to the fact that and the effects of soil evaporation has been found to contribute only 5% of the total when LAI reaches 4 (Shuttleworth and Wallace 1985). Soil evaporation is neglected in the Penman Monteith equation and has not been considered separately in this study. If it is important, better results may be achieved by using a model such as that of Shuttleworth and Wallace (1985) which takes account of evaporation from different surfaces. Herbst *et al.* (1996) calculated transpiration from Penman Monteith equation calculating r_s from porometer measurements, soil evaporation was calculated from lysimeters and total evaporation was calculated using the Bowen ratio approach. Total evaporation from Bowen ratio was found to approximately equal the sum of soil evaporation and transpiration to within 15%. Using all three methods gives added certainty but any two should allow the partitioning of energy between the three sources.

4.5.4 Comparison between top down and bottom up stomatal resistance

When top down and bottom up stomatal resistance values were compared there were mixed results. On 29/08/01 there was a good match but on the three days measured earlier in the season the match was much poorer. Many authors have found discrepancies between bottom up and top down measurements due to the fact that the canopy resistance in the Penman Monteith equation is not purely a physiological parameter (Thom 1975). Alves *et al.* (1998) found that, as in this study, on complete cover crops measured stomatal conductance values were usually lower than those found by the top down method and attributed this to the fact that not all leaves contribute equally to transpiration. Again, another major discrepancy between the r_s of the Penman Monteith equation and measured r_s is that the measured r_s doesn't include other evaporation sources such as soil and intercepted water (Raupach and Finnigan 1988).

4.5.5 Modelling stomatal resistance

Due to the labour intensive nature of taking stomatal conductance measurements, only limited data are available and therefore models are required. To start this process the data were more closely investigated on a daily basis. It was found that the relationship between ET from the Bowen ratio data and from the Penman Monteith equation had a characteristic curved pattern indicating that the Penman Monteith equation was over-predicting when evapotranspiration was low and under predicting when evapotranspiration was high. It was thought an improved relationship could be created if a more realistic changing stomatal resistance could be used. It was found that simply using two different values at different times in the day dramatically improved relationships in the majority of cases. Unfortunately these values were not consistent between different days.

In order to model these changes in stomatal resistance attempts at finding relationships between measured stomatal conductance and environmental factors were made. This is notoriously difficult and it comes as no surprise that the relationships are weak as r_s is simultaneously affected by several variables at once. The relationships are complicated by interactions between the environmental factors, variability in the natural environment, diurnal hysteresis, the long response time of stomata to environmental factors and with amphistomatous leaves like those of *Phragmites* it has been found that stomata on the upper leaf surface tend to be more responsive than those on the lower. Significant correlations were found between stomatal conductance and radiation and vapour pressure deficit. It is well established that stomata respond to light. The proposed mechanism is that light provides energy to drive a potassium-proton pump that alters the turgor pressure in the guard cells surrounding the stomata. In general maximum stomatal conductance is achieved at around 200 W m^{-2} although in this study it was found that conductance continued to increase beyond this figure, possibly due to other factors still being limiting. The role of vapour pressure deficit in stomatal opening is less clear, though it has been found to be influential in many species. There is no consensus however on the physiological basis of the relationship. It may be that the stomata respond to transpiration loss rather than actual humidity (Lhomme 1998). There is a lot of scatter in all the relationships. This may be partly due to the fact that parameters were averaged over 20 minute time periods in which there may be fluctuations in meteorological variables and the stomatal response time may be longer.

Vanyarkho (1996) also attempted to find relationships with environmental variables. Diurnal changes were measured on a single day in August. Single leaf stomatal conductance primarily responded to changes in photosynthetic photon flux density (PPFD) and also to vapour pressure deficit and leaf temperature. Stomatal conductance declined as vapour pressure deficit and leaf temperature increased.

Stomatal resistance proved very difficult to model. A simple regression was unsuccessful. The Jarvis model appeared to have the greatest chance of success. When using top down stomatal resistance data the functions of each parameter are effectively defined by the form of the Penman Monteith equation. Therefore those that were found in this study are the same as those used by Alves and Pereira (2000) and other workers. However there are a large number of parameters that must be defined empirically and it was impossible to do this with any kind of success. The model became extremely sensitive to some parameters so that huge errors developed. In the end it proved impossible to create a working model of stomatal conductance with a consistent parameter set that could be used predictively. The problem of variable parameters was also found by Jarvis (1976) and Wever *et al.* (2002).

4.5.6 Significance of a wet canopy

As in Chapter 3, the issue of the wet and dry canopy once again revealed itself as important. Different curved patterns were seen in the 20 minute data on wet days where Penman Monteith under-predicts for the majority of the time. In most cases this could be improved by setting stomatal resistance to zero. This lends more weight to the hypothesis postulated in the previous chapter that the difference in response on wet days is due to the presence of liquid water on the leaf surface which can be evaporated with zero stomatal resistance. Thom and Oliver (1977) note that when a region is very wet r_s will approximate to zero because evaporation occurs directly from intercepted rain. This means that r_s will be less in a wet month than in a dry one by an amount depending on the fraction of the month when the surface is wet.

4.5.7 Parameterising the Penman Monteith equation for reeds

When the model created using the measured stomatal conductance data was fed into the Penman Monteith equation this resulted in an underestimation of evapotranspiration. This would be expected, as measured stomatal resistance was consistently higher than that predicted by top down Penman Monteith, due to the fact that soil evaporation and interception are not included, and all leaves in the canopy may not contribute equally to transpiration. As it was not possible to create a working model using the top down data the attempt at using a changing value for stomatal conductance had to be abandoned. Best fit values were used. There was a significant difference in these for wet and dry days and there was still a lot of scatter around the 1:1 line. It was found that using this model resulted in a slight improvement over the use of crop coefficients. However in the end both approaches resulted in fairly poor models.

4.5.8 Significance of the coupling factor

The coupling factor is an important parameter in the understanding of reedbed evapotranspiration. There has long been a debate as to the importance of stomata as opposed to energy factors in controlling evapotranspiration (Jarvis and McNaughton 1986). This is particularly pertinent as regards wetland macrophytes, which have often been regarded as passive wicks to water loss. The coupling factor for reedbeds was found on average to be 0.316. This is similar to other reported values for crops around the height of reeds (Jarvis and McNaughton 1986). This indicates that the reedbed is fairly well coupled to the atmosphere with good rates of heat and mass transfer (shown by the low values of aerodynamic resistance) and therefore evapotranspiration is mainly determined by stomatal conductance. This is borne out by the significant relationships found between rates of evapotranspiration and stomatal conductance. The average value 0.684 for $1-\Omega$ indicates that the fractional change in transpiration caused by opening and closing stomata will be $0.684 \times g_s$. This dependence on stomatal conductance is unfortunate however as regards the attempt to parameterise the Penman Monteith equation. A correct description of g_s is important for modelling evapotranspiration. It appears extremely difficult to model stomatal resistance accurately and yet the vegetation is very sensitive to this parameter.

The coupling factor may also give us a clue as to why the Penman Monteith reference equation has been used successfully on short crops such as grass but is more difficult on taller crops such as reed. Ω is strongly dependant on aerodynamic characteristics. A low Ω occurs with higher g_a (lower r_a) above rough and tall vegetation such as reeds. However short grass is typically smooth and poorly coupled to the atmosphere and therefore will be fairly independent of stomatal resistance. Therefore a single value of stomatal resistance is likely to be more realistic in grass.

The influence of intercepted rainfall may also be very important in interpreting the difference in evapotranspiration between tall crops such as reeds and short crops such as grass. Transpiration from tall crops is generally controlled by stomatal conductance and therefore the effective removal of stomatal conductance will mean a large increase in evaporation. In short crops this will make much less difference.

4.6 CONCLUSIONS

Calibrating the Penman Monteith equation directly for reeds is a difficult task due to the complexity of the parameters involved and the many factors that influence them. Ultimately there has been no success in the original aim of creating a successful parameterisation of the Penman Monteith equation for reeds, with only a very slight improvement on the crop coefficient approach. However despite this, significant progress has been made in the understanding of reedbed evapotranspiration.

The aerodynamic resistance of reeds is smaller than for reference grass. Intuitively it would be thought that this would lead to an increase in evapotranspiration but, particularly in 2002, this was found not to be the case and in many cases evapotranspiration was decreased compared to reference evapotranspiration with a lower aerodynamic resistance whilst all other variables were held constant. This may be part of the explanation as to why reed evapotranspiration was found to be lower that reference grass.

The coupling results indicate agreement with Sanchez-Carrillo *et al.* (2001) who state that reed transpiration is controlled by stomata rather than aerodynamic conditions. This means that it is very important to be able to accurately predict stomatal resistance. Although it could be measured and seen to have some relationship to environmental variables the sum of measured stomatal conductance was not equal to the surface resistance parameter of the Penman Monteith equation. This may be due to significant soil evaporation adding to the overall evapotranspiration and due to the fact that the parameters are not necessarily physically equal. Again the important impacts of interception were seen and this is an important factor to take account of in modelling. "Especially where tall vegetation predominates, therefore, attempts to unravel the water balance of a vegetated region will be unrewarding unless there is adequate recognition of the effects of surface roughness and surface wetness" (Thom and Oliver 1977 p.356).

The lower values of measured stomatal conductance compared to reference evapotranspiration (Chapter 3) appear to be explained by firstly a lower aerodynamic resistance and secondly a higher stomatal resistance. However this may not actually correspond to a higher surface resistance, as this is reduced by the influence of soil evaporation.

CHAPTER 5 - The Water Balance of Stodmarsh National Nature Reserve

5.1 INTRODUCTION

5.1.1 Why create a water balance?

"Hydrology is probably the single most important determinant of the establishment and maintenance of specific types of wetlands and wetland processes" (Mitsch and Gosselink 2000 p.108). All natural wetland functions are the result of, or are closely related to, hydrology (Carter *et al.* 1978). It is therefore vital for wetland scientists to have an understanding of wetland hydrology for management, conservation and creation. A water balance is a common tool used in the hydrological analysis of wetlands (Carter 1986) and can be regarded as a prerequisite for fully understanding wetland hydrology and the movements of water into and out of the wetland (Kadlec 1990). Water balances provide essential support for management decisions, allow assessment of the water requirements of habitats and are necessary for the calculation of nutrient budgets and energy fluxes and for the prediction of the effects of change. However, despite the importance of studying the water balance, "solid quantitative information about the hydrodynamic characteristics of different wetlands is surprisingly difficult to find" (Gosselink and Turner 1978 p.64). The complex nature of wetland-watershed relationships mean that there is still a great deal of uncertainty about the water balance of wetlands and most existing studies have been done on fairly simple catchments (Owen 1995). There is therefore a need for more studies of this type to increase understanding of wetland hydrology.

5.1.2 What is the water balance of a wetland?

The main hydrological fluxes of a reedbed are shown in the schematic diagram (Figure 5.01).

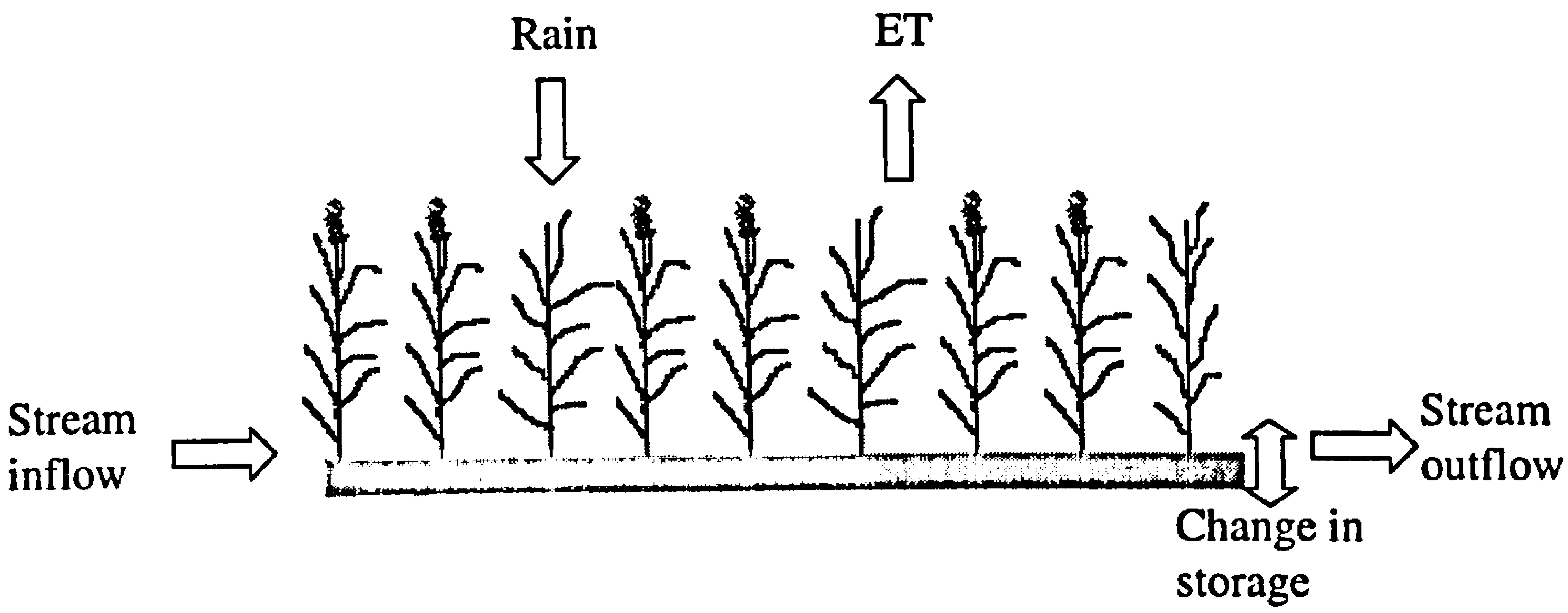


Figure 5.01 – Schematic diagram of the major hydrological fluxes in a reedbed.

The general water balance of a wetland is:

$b = P_n + S_I + G_I - ET - S_o - G_o + \Delta V/\Delta t$ (Dooge 1975) (5.01)

where b is measurement error, $\Delta V/\Delta t$ is the change in volume of water stored in a wetland per unit time, P_n is precipitation, ET is evapotranspiration, S_I is surface inflow, S_o is surface outflow, G_I is ground water inflow and G_o is ground water outflow. Ideally b should be as small as possible.

A complete water balance is composed of independent and reliable measurements of each component (Dooge 1975). If all components are measured the deviation of the water balance from zero can be calculated. This allows validation of the water balance model and an assessment of the reliability of the measurements. However it is rare that all components are included due to limitations on data collection. Unmeasured components of the water balance are often calculated as residuals but inevitably this will include all the error from the other terms and so has some degree of uncertainty. If the water balance equation can be balanced and the deviation is close to zero it is likely that the measurements or estimations of the water balance are satisfactory and that the stream catchment comprises a watertight hydrological unit. If the equation cannot be solved it is necessary to investigate whether certain components are consistently under or over estimated.

5.1.3 Hydrologic wetland classification

There are numerous hydrogeomorphic classification systems for wetlands (e.g. Cowardin *et al.* 1979; Brinson 1993a). Classification can be based on the amount of surface water, nutrient inflow, vegetation type, pH and peat building characteristics. All of these features of wetlands are ultimately rooted in hydrology and therefore hydrology seems the logical base for a functional classification (Keddy 2000).

Water creates wetlands and is the key determinant of a vast array of wetland processes and properties, and these in turn result in a wide range of wetland types. Differences in plant communities between bogs and fens, for example, are due to nutrient availability, which is in turn a function of hydrology. There are three main sources of water for wetlands – precipitation, surface flow and groundwater. Infertile, unproductive bogs are dependent on rainfall (ombrotrophic) and isolated from surface flow and therefore have little opportunity to influence the quality of ground and surface water. Fens are connected to flowing groundwater (minerotrophic) and depend on aquifer discharge. Riverine wetlands may also be dominated by groundwater or by overland flow which includes over-bank flooding (Brinson 1993b) and seepage to and from a river, infiltration of floodwater and overland flows (Bradley 2002). It is fairly common for wetlands to be little influenced by ground water as they often occur where soils have poor permeability and the major loss of water is evapotranspiration and surface outflow (Mitsch and Gosselink 2000). Wetlands can be classified using the water balance according to the relative proportions of the three sources of water making up their inflow using the triangle diagram in Figure 5.02 (Brinson 1993a; 1993b).

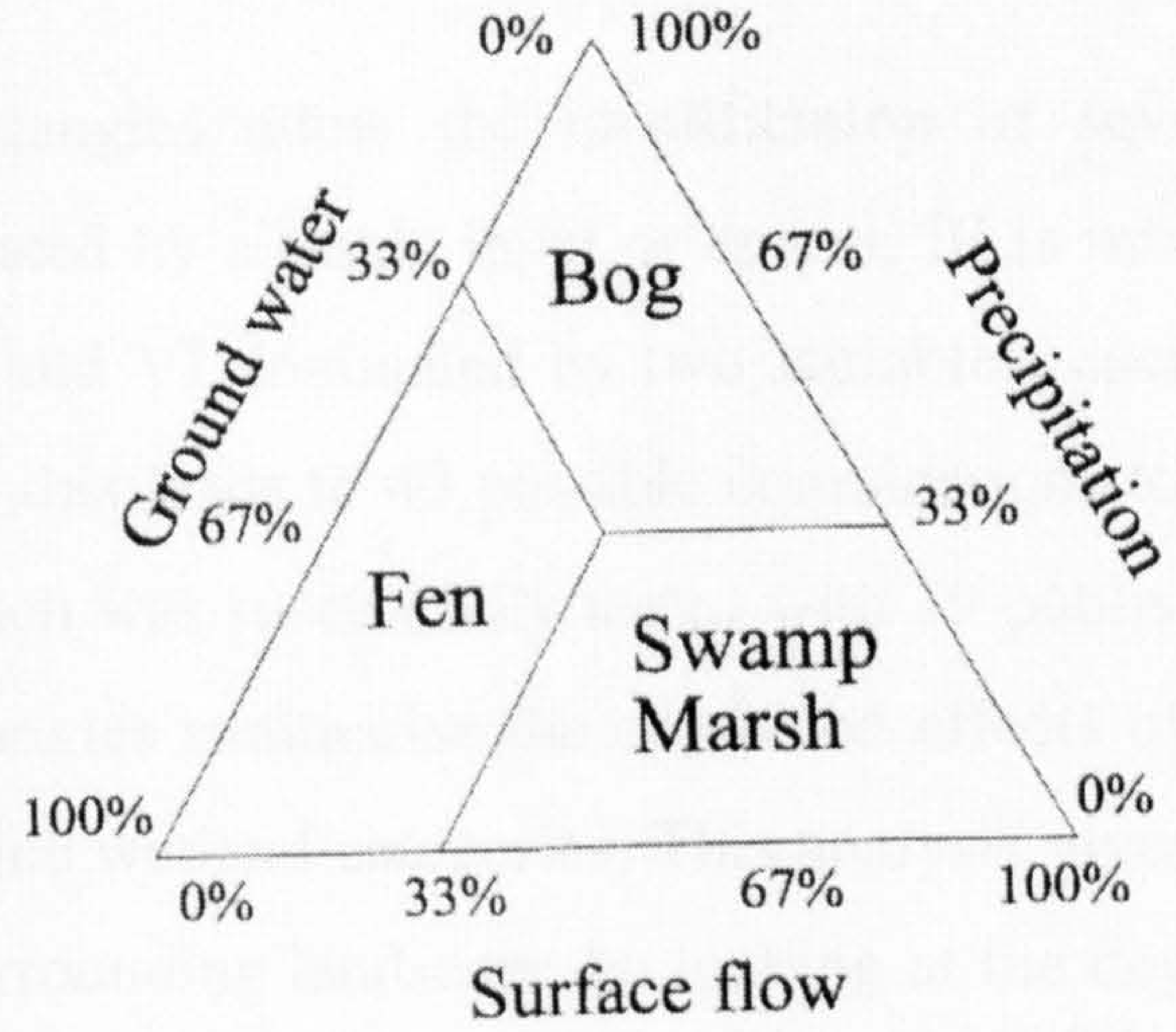


Figure 5.02 – Classification of wetlands based on the relative importance of water inputs (from Brinson 1993b).

Lent *et al.* (1997) provide a more complex, quantitative classification system for wetlands based on the degree of groundwater input, surface water input and precipitation inputs and also includes outputs. These are plotted on tri-linear diagrams and then inputs and outputs are summarised in quadrangle shaped diagrams as can be seen in Figure 5.03.

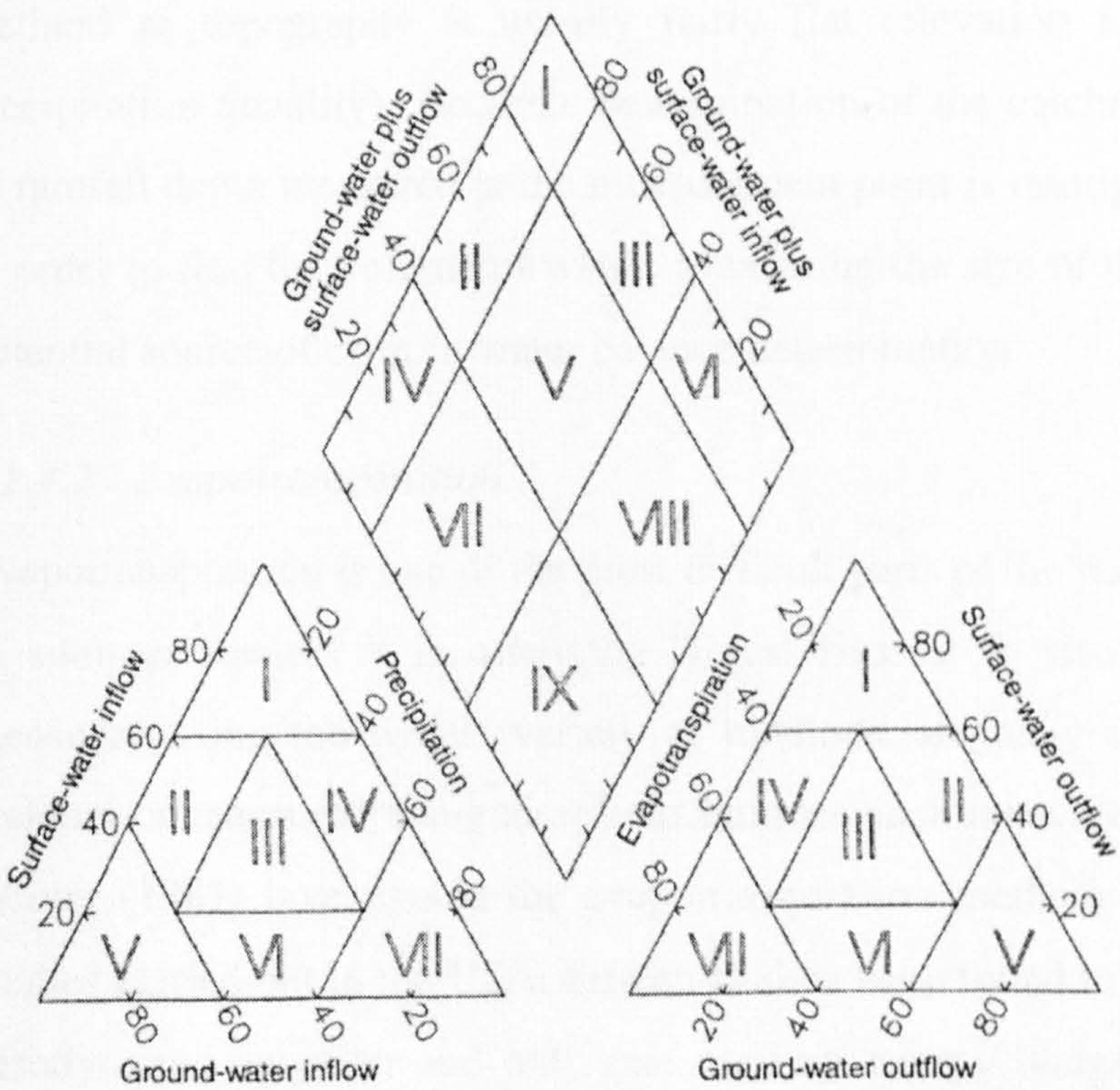


Figure 5.03 – Classification of wetlands based on the water balance (Lent et al. 1997)

The triangles allow the identification of seven wetland types: I, V and VII are dominated by a single input or output, III is mixed with all three greater than 20% and II, IV and VI dominated by two variables, each less than 60%. Using both input and output this leads to 49 possible combinations to describe the hydrological regime. The approach was successfully tested with 19 published wetland water balance studies. The quadrangles summarise the combined effects of inputs and outputs. These are divided into nine wetland categories. This analysis gives information about the interaction with the surrounding landscape by looking at the degree of domination of the water balance by precipitation and evapotranspiration.

5.1.4 Water balance components

5.1.4.1 Precipitation

Precipitation has an immediate and visible effect on the water conditions in wetlands and in the majority of wetlands is the dominant water source. Precipitation quantities are determined from a small point source which is assumed to be representative of the surrounding area. Fortunately, spatial variation of rainfall is often small within a wetland as topography is usually fairly flat (elevation is a major determinant of precipitation quantity). Accurate determination of the catchment size is also important as rainfall depth measured at the measurement point is multiplied by the catchment area in order to find the volume of water. Measuring the size of the catchment is therefore a potential source of error in water balance determination.

5.1.4.2 Evapotranspiration

Evapotranspiration is one of the most difficult parts of the water balance to measure, but in summer months it is often the largest flux. It is also probably the flux that is measured using the widest variety of methods. In many studies it is estimated as a residual or measured using simple techniques such as evaporation pans (Carter 1986). Winter (1981) investigated the evapotranspiration methodologies of 23 water balance studies carried out in the USA. Fifteen studies were found to use pan data, five to use an aerodynamic equation and only one used an energy budget technique. However this latter method was considered to be the most accurate technique for periods of a week or longer. Pan data have a large number of problems including advection effects, the influence of the design of the pan, and most seriously, in choosing coefficients to relate the evaporation of the pan to that of the habitat for which the water balance is being created. Other methods have also been used to estimate evapotranspiration. Rushton (1996) estimated evapotranspiration by observing diurnal changes in water level, Koerselman (1989) used lysimeters to calibrate a model based on Penman's open water equation and Rouse (1998) used the Bowen ratio energy balance method.

Evapotranspiration models are also commonly used. Drexler (1999) used MORECS (UK Meteorological Office Rainfall and Evaporation Calculation System) and Bradley (1996) used the Priestly-Taylor equation. Lott and Hunt (2001) compared the Penman combination equation used with a wind function with measurements of evapotranspiration from lysimeters and from fluctuations in the water table. They found the calculated evapotranspiration was lower than that which was measured. They concluded with the recommendation that that there may be other meteorological equations which might better represent evapotranspiration from a wetland in future studies. They suggested the use of Penman Monteith, as it includes parameters of canopy structure and roughness that may provide better representation of evapotranspiration, by reflecting differences in advective transport capacity governed by the differences in windspeed. The use of the Penman Monteith equation with crop coefficients is a common general evapotranspiration modelling technique (Pereira *et al.* 1999), though this has not been widely used in natural environments such as wetlands. A crop coefficient is the ratio of crop evapotranspiration (ET_c) to reference evapotranspiration (ET_{ref}). ET_c is evapotranspiration from a specified vegetative surface in good agronomic condition, extensive in area, under optimum soil water conditions and achieving full production under the given climatic conditions. ET_{ref} was defined as the evapotranspiration of a surface with a height of 0.12 m and bulk surface resistance of 70 s m^{-1}, with leaf area index similar to short grass as estimated using the Penman Monteith equation (Allen *et al.* 1994c). ET_{ref} is only a function of the weather. Once coefficients have been developed they can be used to estimate ET_c using Equation 5.02.

$$ET_c = K_c ET_{ref} \tag{5.02}$$

The advantages of the Penman Monteith equation are that it is physically realistic and that it only requires standard meteorological measurements as inputs. However the use of other, simpler, equations has also been attempted.

5.1.4.3 Surface flow

Flow in streams is measured using structures such as weirs or velocity-areas methods such as Doppler flow meters. The greatest problem is in estimating surface flow occurring as sheet flow over the surface adjacent to the wetland (Carter *et al.* 1978). It is often omitted due to measurement difficulties and assumed to be negligible (Koerselman 1989; Drexler *et al.* 1999).

5.1.4.4 Groundwater flow

Groundwater is one of the hardest components of the water balance to measure (Bradley 1996). The hydraulic gradient is usually investigated with observation wells, piezometers and dipwells and fluctuates seasonally. In order to make accurate estimates of fluxes it is necessary to have an understanding of soil properties, in particular hydraulic conductivity, which is hard to ascertain and can be variable over the site.

5.1.4.5 Input from rivers

River input can be a significant component of the water balance in floodplain wetlands. However its magnitude is difficult to determine and there are few examples of measurements. A piezometer network can allow a qualitative assessment of subsurface flow direction. Over-bank floods can also be important but quantification would require measurement of spatial variation in infiltration, hydraulic head, river stage and depression storage (Bradley 1996).

5.1.4.6 Storage

Stored water can either be ponded water on the surface of a wetland, or held within the soil. When the wetland is unsaturated, it is monitored by measuring water table change taking account of soil porosity whilst when there is ponded water depths can be recorded. Over long periods of time storage change is often assumed to be zero. Wetland hydrology often places emphasis on storage, as it is the parameter that wetland managers are most interested in. Water levels in open water areas are susceptible to changes due to manipulation by land managers as well as seasonal changes and changes due to growth, cut-back and die-back of vegetation (Gilman 1992).

5.1.5 Examples of water balances

Despite the importance of wetland hydrology few studies have quantified inflows or total wetland water balances, especially in the UK (Gilvear *et al.* 1993). LaBaugh (1986) carried out a review of wetland water balance studies which give the most complete information on wetland water balances up to the time of writing. Few studies attempted to measure most parts of the water balance. Examples cited included Crisp (1966) who studied peat bogs in England, Mitsch *et al.* (1979) who investigated alluvial swamps in Illinois, Verry and Timmons (1982) who studied upland peat in Minnesota, Hemmond (1980) who worked on floating mat Sphagnum bog and Eisenlohr (1975) who worked on prairie potholes. All used different hydrological methods and, often, particular components were ignored because they were considered negligible or were calculated as a residual. The ideal of measuring all fluxes was rarely achieved due to data limitations.

A rare example of a complete water balance is the work of Koerselman (1989) who created a water balance of a small groundwater fed fen. Observation wells and piezometers were installed in order to monitor groundwater change and staff gauges were installed in ditches. Precipitation was determined using raingauges and evapotranspiration measured using lysimeters, which were used to calibrate a regression model against Penman potential evaporation.

A major UK study is that of Gilman and Newson (1983) who created a water balance as part of the Anglesey Wetland study. They found that more than half the rainfall input to the Anglesey wetlands is lost as evapotranspiration and 80% of annual evapotranspiration occurs between April and September. Ward (1967; 1972) also completed a UK water balance study successfully in Holderness, Kent, using a simple water balance approach, using simply streamflow, precipitation and evapotranspiration, though a change in storage component was not included. The site had a heavy clay soil and the researchers assumed the catchment was watertight. A water balance was created over three years, one year and over individual seasons and months. It was found that the balance was good over three years but over the shorter periods change in storage was important. They also checked the water balance by predicting components of the balance from the residual of the water balance and comparing these with measured data.

Gilvear *et al.* (1993) created a water balance on Badley Moor, East Anglia, which is a groundwater fed wetland. Rainfall and evapotranspiration data were obtained from MORECS and the wetland was modelled using MODFLOW, a groundwater simulation model. Groundwater accounted for 90% of the inputs and surface water output for 87% of the outputs. Measurements were made at monthly intervals due to the slow changes in the groundwater system.

Bradley (2002) created a water balance on Narborough Bog, Leicestershire, UK, which includes areas of *Phragmites,* alder carr and wet meadow. Over-bank floods occurred at regular intervals. The research aimed to reproduce the temporal pattern of the water level regime for a floodplain wetland. They had a two-year measurement programme, which included measurements of wetland water tables using piezometers, the river stage and meteorological parameters.

Rushton (1996) attempted to quantify a water balance for a freshwater marsh in Florida. Evapotranspiration and groundwater seepage were considered to be the most difficult components of the water balance to measure. Evapotranspiration was estimated by observing diurnal changes in water level and inflow and outflow were measured using weirs.

Drexler *et al.* (1999) determined the water budget of a small peatland in New York State, USA. Groundwater flux was estimated using a chemical mass balance method with Fick's Law, precipitation with a raingauge, streamflow with a weir and evapotranspiration was estimated using MORECS. Groundwater was found to be the main flow onto the reserve and precipitation was quite low.

5.2 OBJECTIVE

The objective of this chapter is to create a water balance of Stodmarsh National Nature Reserve that is as complete as possible, using the most accurate measurement methods available. This will be valuable research in itself given the small number of complete water balance studies carried out in the UK. The validation of the completed model will show whether the water balance model being employed on the site is appropriate, whether all inflows and outflows are being accounted for and if the measurements are sufficiently accurate. Residuals of the water balance will be used to create additional estimates of evapotranspiration and predictions of the change in storage on the site. Models will subsequently be created from the resultant water balance in Chapter 6.

5.3 MATERIALS AND METHODS

5.3.1 Study site

The experimental work was carried out at Stodmarsh National Nature Reserve, Kent, UK (51° 19'N, 1°12'E, elevation <5 m). The site has fairly simple hydrology. It is underlain by London clay over Woolwich and Thanet beds, which has a very low permeability and therefore groundwater percolation is considered insignificant. Inputs are from the Lampen Stream and precipitation and the outputs of water are stream outflow and evapotranspiration. A diagram of the site can be seen in Figure 5.04, to which the numbers below refer. The features of the site are:

The Lampen Wall (1), a clay bund, divides the SSSI into the eastern (2) and western (3) areas. The western area contains a large lake and coal spoil heap, but as it is not part of the NNR this area is not included the water balance. The site is bordered by the River Great Stour (4) on the north-western and eastern sides but the two systems are in theory hydrologically distinct being divided by the continuation of the Lampen Wall. The Lampen Stream (5) flows into the eastern area, controlled by a sluice (6). A system of sluices and dams allows the water levels in the site to be controlled in three distinct areas. To the east, a road borders the site (7), and only 0.38 km away is the River Little Stour (8), so the catchment is small and therefore non-channelised surface flow is limited. There is, however, the possibility of water seeping through the clay bund from the river, some evidence of which has been seen at times of high river flow. The site has an area of established reedbeds (referred to as the "old area") and in 1995 an additional 79 ha of fields adjoining (and now part of) the NNR were purchased and planted as reedbeds to increase the area of the site (referred to as the "new area").

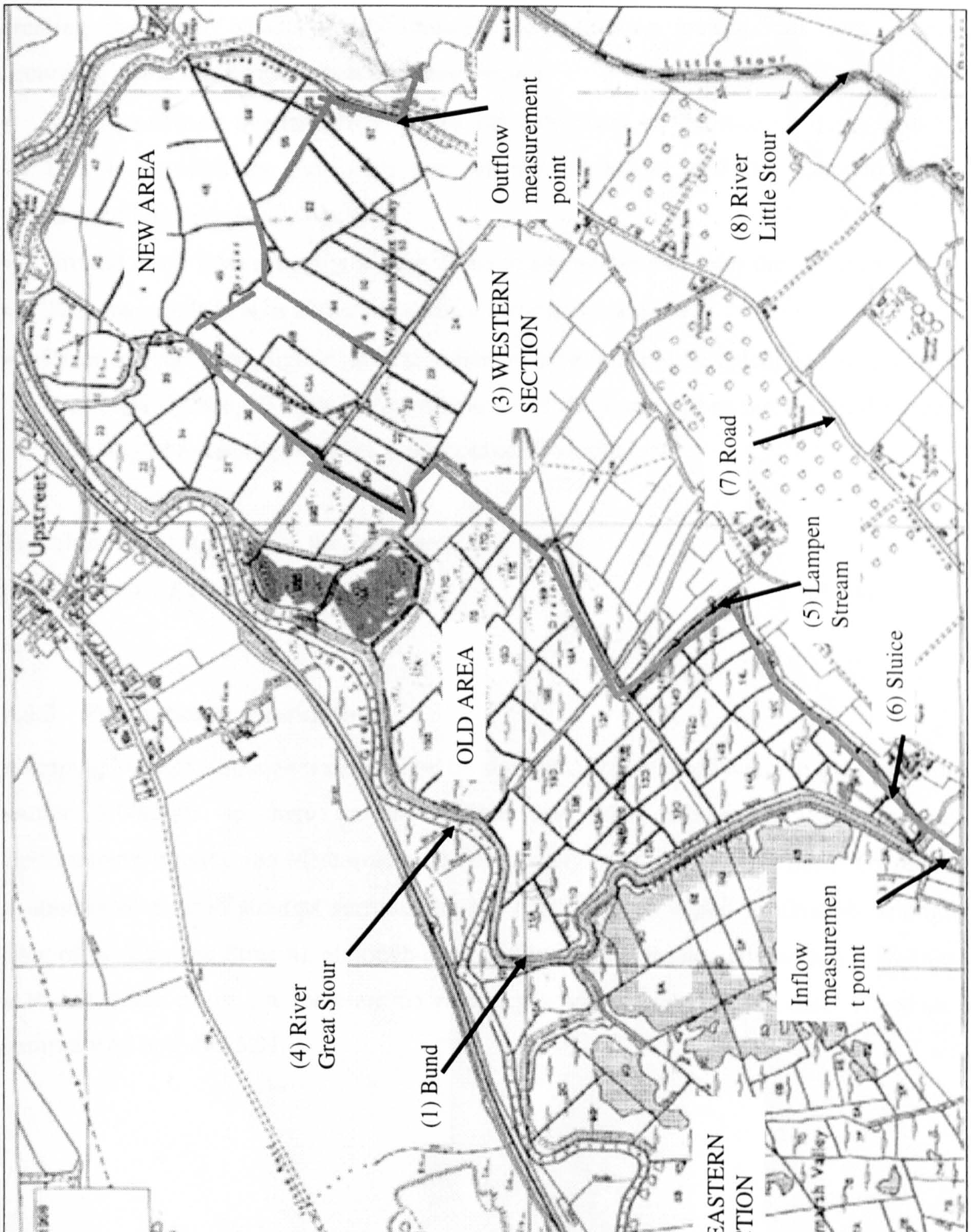


Figure 5.04 – Site map of Stodmarsh National Nature reserve showing the main hydrological features.

5.3.2 The water balance of Stodmarsh NNR

It is important to be clear about the assumptions of the study (Dooge 1975). When creating the water balance for Stodmarsh the following assumptions were made according to the information in the site description:

- Non-channelised overland flow across the catchment to the reserve is negligible. This is justified by the small size of the catchment and the Lampen Wall surrounding two sides of the site.
- Ground water flow is negligible due to the impermeable nature of the substrate.
- The water table is at or above the surface throughout the site.
- River seepage is negligible and the water balance will not be calculated at times of over-bank flooding. This is the most tenuous of the assumptions but due to the great difficulty in measuring seepage it is necessary to include it.

Therefore the water balance for Stodmarsh is:

$$b = P_n + S_I + - ET - S_o - + \Delta V/\Delta t \qquad (5.03)$$

5.3.3 Precipitation measurement

A tipping bucket raingauge was installed on site but it was washed away in the floods of winter 2000/2001 so there are no reliable data from Stodmarsh itself. Instead Environment Agency and Meteorological Office daily rainfall data were used. There are a number of rainfall stations surrounding the catchment to which Stodmarsh belongs (that of the Lampen Stream), although there are no raingauges actually within it. Within a 10 km radius of the site there are six raingauges with current rainfall data. These are summarised in Table 5.01.

Table 5.01 – Details of raingauges within 10 km of Stodmarsh

Station	Latitude and Longitude	Elevation	Distance from Stodmarsh
West Stourmouth	51.320°N, 1.229°E	9 m	1 km
Sturry	51.18°N, 1.07°E	21 m	4.5 km
Canterbury	51.252°N, 1.216°E	5 m	6 km
Herne Bay	51.356°N, 1.134°E	40 m	6 km
Garrington	51.265°N, 1.158°E	12 m	5.5 km
Bekesbourne	51.15°N, 1.08°E	15 m	6.5 km

Thiessen polygons were used to estimate the relative importance of each of these stations in determining the rainfall at Stodmarsh. This is an objective technique which sums rainfall measurements made at each gauge, weighted by the fraction of the catchment area represented by the gauge. Areal rainfall is:

$$P_n = \sum \frac{P_{ni} a_i}{A} \tag{5.04}$$

where P_{ni} is rainfall measurements at n gauges, a_l is polygon area corresponding to raingauges and A is catchment area. The catchment was divided into polygons with lines equidistant between stations. The area of each segment was measured using squared paper. The polygons are illustrated in the schematic diagram in Figure 5.05.

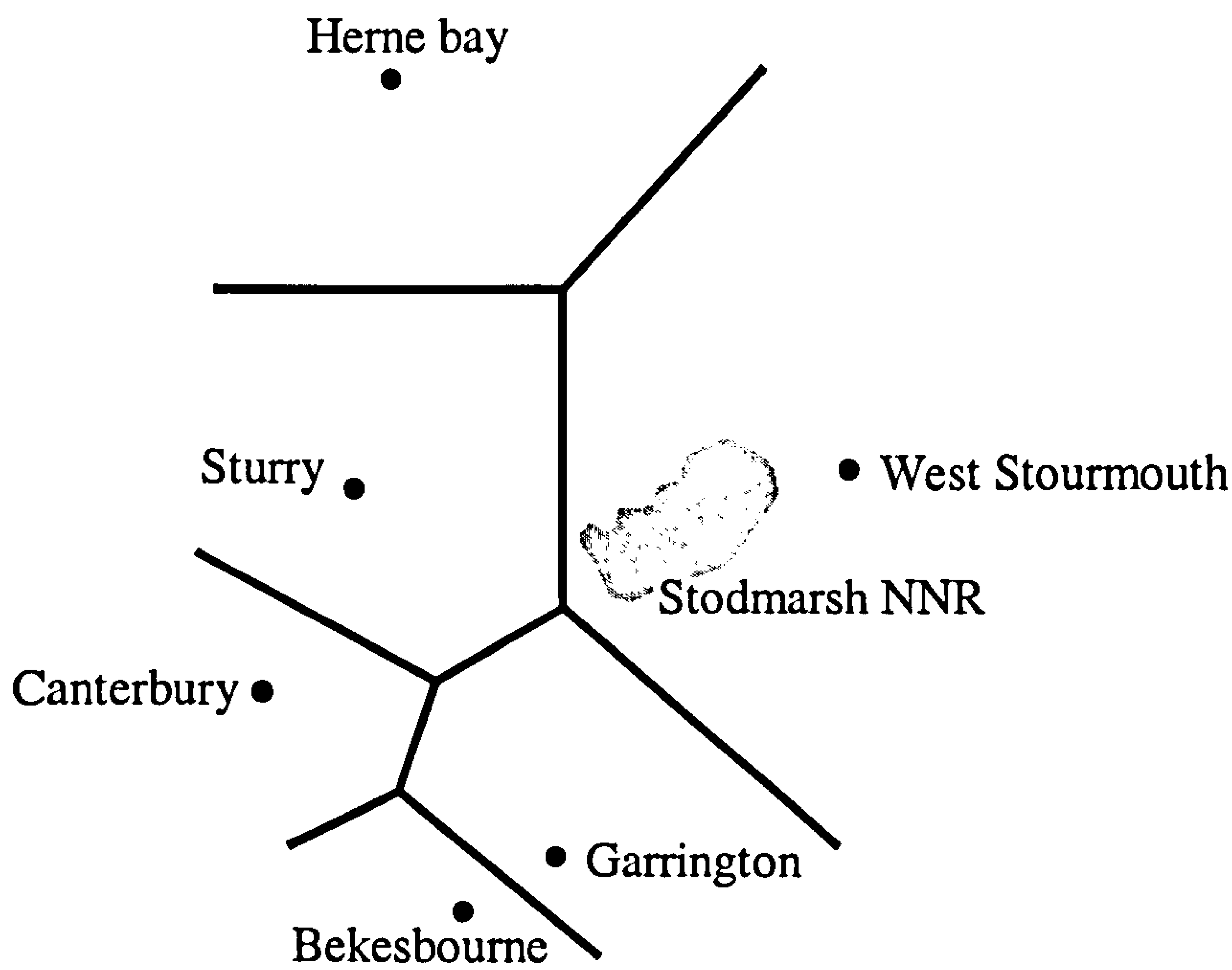


Figure 5.05 – Schematic diagram of the location of raingauges (•) surrounding Stodmarsh NNR and the Theissen polygons drawn around them (bold lines) (not to scale).

The daily rainfall data were converted from mm d^{-1} to m^3 by multiplying by the site area (242.9 ha).

5.3.4 Stream flow

Discharge at the inflow and outflow points of the Lampen Stream from the reserve was measured using ultra sonic Doppler meters (Starflow, Unidata Ltd.). These have been shown to be an accurate method of measuring streamflow (ISO technical report 2000). They measure mean channel velocity by transmitting a high frequency sound wave, which is reflected back by particles moving in the water and a frequency shift is detected. This shift is proportional to the velocity of the moving particle and the velocity can be calculated from the magnitude of the Doppler shift, the angle between the transmitted beam and direction of movement and the speed of sound in water. The mean of a large number of particles is found. Velocity measurement is thought to be accurate to within ±2%. Stage is measured using a pressure transducer (accuracy ±0.25%).

The inflow to the site was measured just outside Stodmarsh village within a large rectangular culvert under the road. The water on installation was only around 0.21 m deep so it was possible to secure the meter to the bed of the stream. The width of the channel is 1.35 m. The outflow measurement site was in a 0.76 m diameter cylindrical pipe and the meter's mounting clamp was held in place by an expanding steel band as drilling into the pipe was impossible.

Based on the geometry of the channel, the cross-sectional area can be established from the measured stage. This was more complex at the outflow point as measurements were made in a cylindrical pipe so that width varies with depth. By multiplying the cross sectional area by the mean velocity the stream discharge can be found. The dataloggers within the Starflow took a reading of velocity and depth every minute and then this was averaged every 20 minutes. These data were used to calculate discharge in m^3 s^{-1} and this was averaged on a daily basis. The data were downloaded every month between April 2000 and October 2002. However due to instrument failure, caused by flooding in spring 2001 and further instrument failure in spring 2002 some data were lost resulting in the water balance not being completed at these times.

5.3.5 Change in storage

The depth of water ponded on the surface of the reserve was measured in order to estimate the volume of water stored on the site. Measurements were made on six occasions between August 2000 and August 2001. The surface water depth measurements were made with a ruler. A global positioning system (GPS) (Trimble, Pathfinder) was used to record the location of the measurements and plot them onto a digitised map. The use of the GPS meant that measurements could be accurately (within less than 1 m) located on the site. The site is large and accessibility is limited so only selected areas were sampled. Five or six compartments (areas of the site bounded by ditches) were sampled on each occasion. The new area of the site is divided into three independently flooded areas and it was ensured that samples were taken from each of these. It is assumed that within these three areas the water finds its own level and is consistent above a datum. Measurements were made every 25 paces, whilst wading through the water. On subsequent sampling occasions it was hoped that measurements would be taken in similar places but accuracy in this proved impossible so the GPS was invaluable in data analysis.

A levelling survey was carried out on the new area of the site by English Nature prior to the planting of the reeds and the commencement of this research. The results of this survey have been placed on a digitised map. The measured depths were plotted onto the same map using the software ArcView GIS (Environmental Systems Research Institute Inc. version 3.2a). This allowed the measured depths to be compared to levelling points close by. The sum of the level above datum and the measured water depth above it were presumed to be approximately equal across each flooded area and this is equivalent to the water elevation above the levelling datum. An average was found of the elevations above mean levelling datum across each flooded area. Any values that were not within two standard deviations of the mean were removed, as it was possible to get incorrect values if there were small ridges or troughs in the bed.

There was no significant difference in elevation above datum between compartments that were flooded separately compared to those flooded together so all the compartments were simply averaged to find an overall average elevation above datum for the new area of the site. Depths measured in close proximity to one another on separate site visits were also compared and the average change in depth for each field and for the whole area was found. These changes were then compared to the changes calculated from the levels.

The change in depth was converted to a change in volume across the site. This was done by assuming each levelling point represented an equal area of the site (they were fairly evenly spread across the site – see Figure 5.11). The depth of water at each point was calculated by subtracting the level at each point from the average elevation above Ordnance datum. If the resultant water depth was negative then the water depth was assumed to be 0.0 m. The average water depth was then found of all these points. This average depth was then multiplied by the area of the new area and converted into a volume (m^3).

A major limitation of the volume calculation was that there is no levelling survey for the old area of the reserve as it is very difficult to survey once the reeds are in place. Because of this it must simply be assumed that the average level on the old area is the same as that on the new and a calculation for the old area was performed in the same way as above.

Gaugeboard readings at the turf fields main stop, close to the outflow were also recorded at the same time as water depth measurements were made. The recordings were related to the volume of water on site and can be used to predict storage volumes for times when there are gaugeboard readings but no depth measurements.

5.3.6 Evapotranspiration

Evapotranspiration was calculated using the Penman Monteith reference equation multiplied by crop coefficients. Weather data from the nearest meteorological station at Manston Airport (51.349°N, 1.353°E, elevation 44 m) around 10 km from Stodmarsh were used to create reference evapotranspiration. The data from Manston were compared to those measured using on site meteorological instruments at Stodmarsh to investigate whether there are significant differences in the climate of the two sites, caused by their different elevations and proximity to the coast.

The average monthly crop coefficients for Walton Lake, Milton Keynes, Buckinghamshire calculated by Fermor *et al.* (2001) (Kc_{Fermor}) were used in the first instance. Although Fermor *et al.* calculated crop coefficients for three sites, Walton Lake is the most similar to Stodmarsh. It is a well-established and extensive site, whereas the other two are small and newly established. The Walton Lake crop coefficients can be seen in Table 5.02.

Table 5.02 – Mean crop coefficients from Walton Lake, Buckinghamshire, UK (after Fermor et al. 2001)

	Jan	Feb	March	April	May	June
Mean *Kc*	1.09	0.46	0.48	0.78	0.63	0.77
S.E.	0.32	0.09	0.09	0.19	0.07	0.07
	July	**August**	**Sept**	**Oct**	**Nov**	**Dec**
Mean *Kc*	0.86	0.72	0.75	0.82	0.76	0.97
S.E.	0.13	0.15	0.02	0.13	0.10	0.57

The Bowen Ratio Energy Balance Technique (BREB) was also used to measure evapotranspiration over reeds at Stodmarsh (see Chapter 3). The results of this were also used in the water balance and compared with the Fermor coefficients in order to validate whether these coefficients are applicable to Stodmarsh.

Crop coefficients were also calculated as residuals from the water balance.

5.3.7 Overall water balance

The components of the water balance were brought together in three ways. Firstly solving the water balance equation as:

$$\Delta V/\Delta t = (P_n + S_I) - ((ET_{ref}\ x\ Kc_{Fermor}) + S_o) \qquad (5.05)$$

The water balance was created on a daily basis, the total volume of each component being calculated over this period allowing a change in storage volume on the site to be estimated. This change in storage estimation was then validated using the measured storage data.

Secondly the balance was calculated using measured evapotranspiration data (from the Bowen ratio approach) and measured change in storage data in order to investigate the magnitude of the discrepancy between input and output using the equation:

$$b = P_n + S_I - ET_{reed} - S_o + \Delta V/\Delta t \qquad (5.06)$$

Finally the equation is rearranged so that the crop coefficients are calculated as the residual so that these can be compared with Kc_{Fermor} and the crop coefficients created by the Bowen ratio technique.

$$Kc = \frac{P + S_i - S_o - \Delta V / \Delta t}{ET_{ref}} \qquad (5.07)$$

This equation was used over periods for which we have reliable storage data.

5.4 RESULTS

5.4.1 Rainfall

Thiessen polygon analysis revealed that the entire reserve fell within the West Stourmouth polygon and therefore these data were used exclusively for calculating the water balance. However this station was no longer recorded after October 2001 and therefore for the final 3 months of 2001 (and in Figure 5.06 below) data from Garrington were used as this is the data set most similar to West Stourmouth (r^2 0.71). Figure 5.06 gives a summary of the rainfall and temperature data during the period for which the water balance was measured.

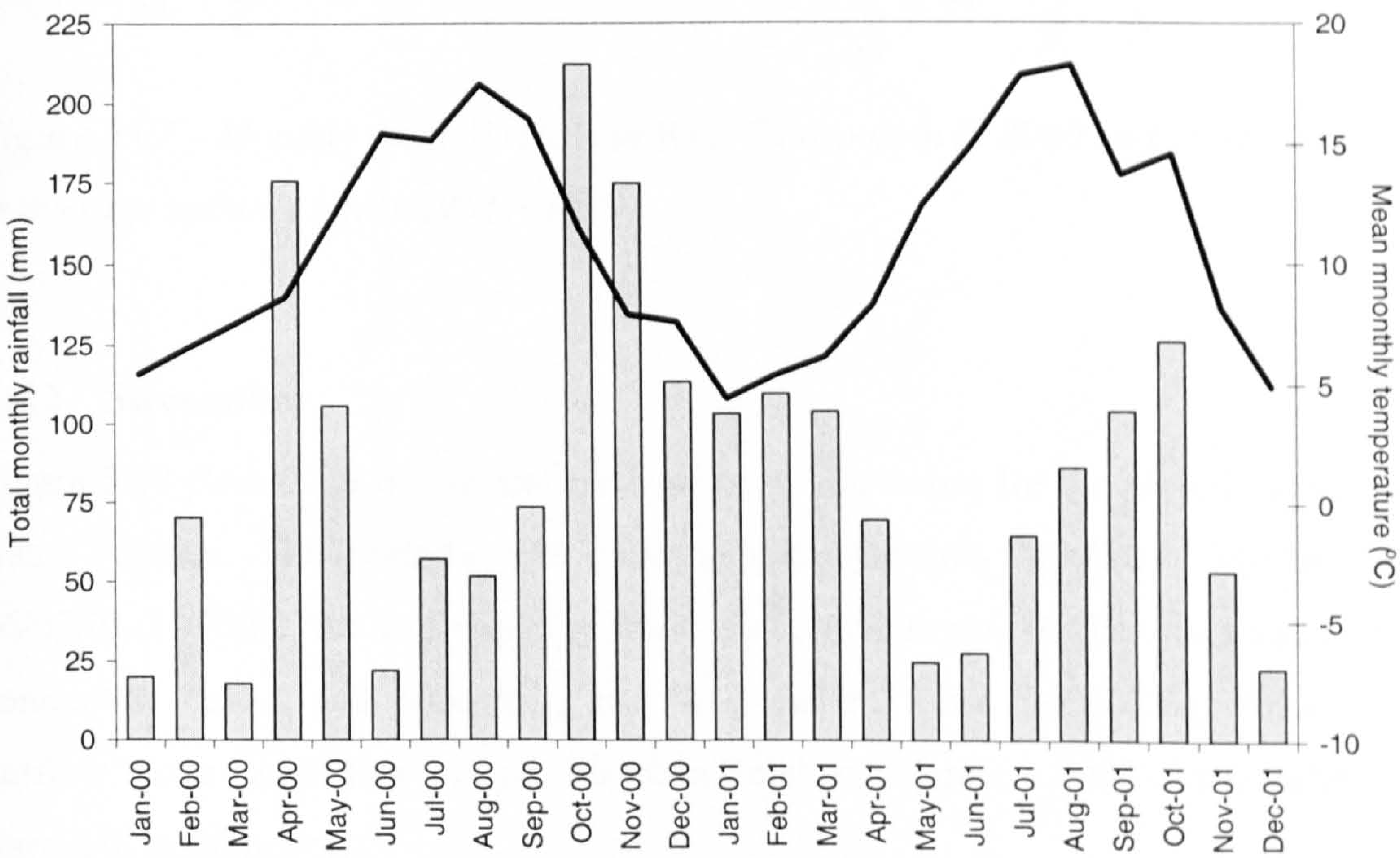


Figure 5.06 – Total monthly rainfall from West Stourmouth (grey bars) and monthly mean temperature data at Manston (black line) for the period January 2000 to December 2001.

Both 2000 and 2001 were particularly wet years with total annual rainfalls of 882 mm and 823 mm respectively. The average annual rainfall of this region for 1964 to 1999 was 611 mm. High rainfall during the winter of 2000/2001 led to data loss due to flooding. However the above average rainfall was not evenly spread throughout the year (Figure 5.07). In 2000 high rainfall occurred in April, May and again in October, November and December. This high rainfall continued until March 2001 and rainfall was also high in August and September 2001.

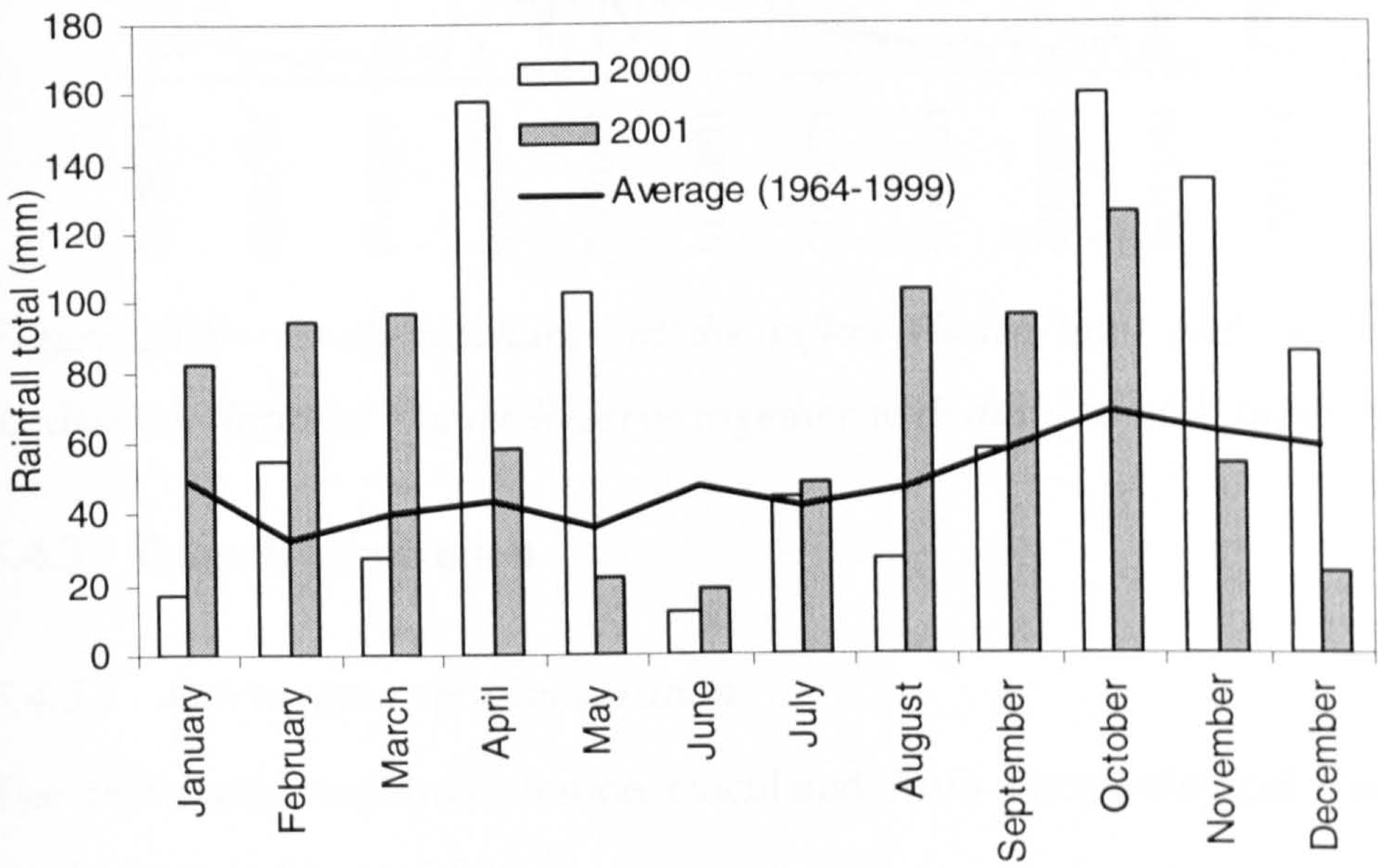


Figure 5.07 – Monthly rainfall totals at West Stourmouth in 2000 and 2001 compared to average monthly totals 1964 – 1999.

5.4.2 Streamflow

Figure 5.08 plots daily inflow and outflow streamflow data for the period used to create water balance. The periods with missing data (outflow 06/02/01-15/05/01, inflow 06/02/01-21/03/01) are explained by flood water submerging the batteries and computer connection cables and preventing the flow meters from functioning. After this the outflow meter had to be replaced. The data finishes in January 2002 as the inflow meter started to malfunction.

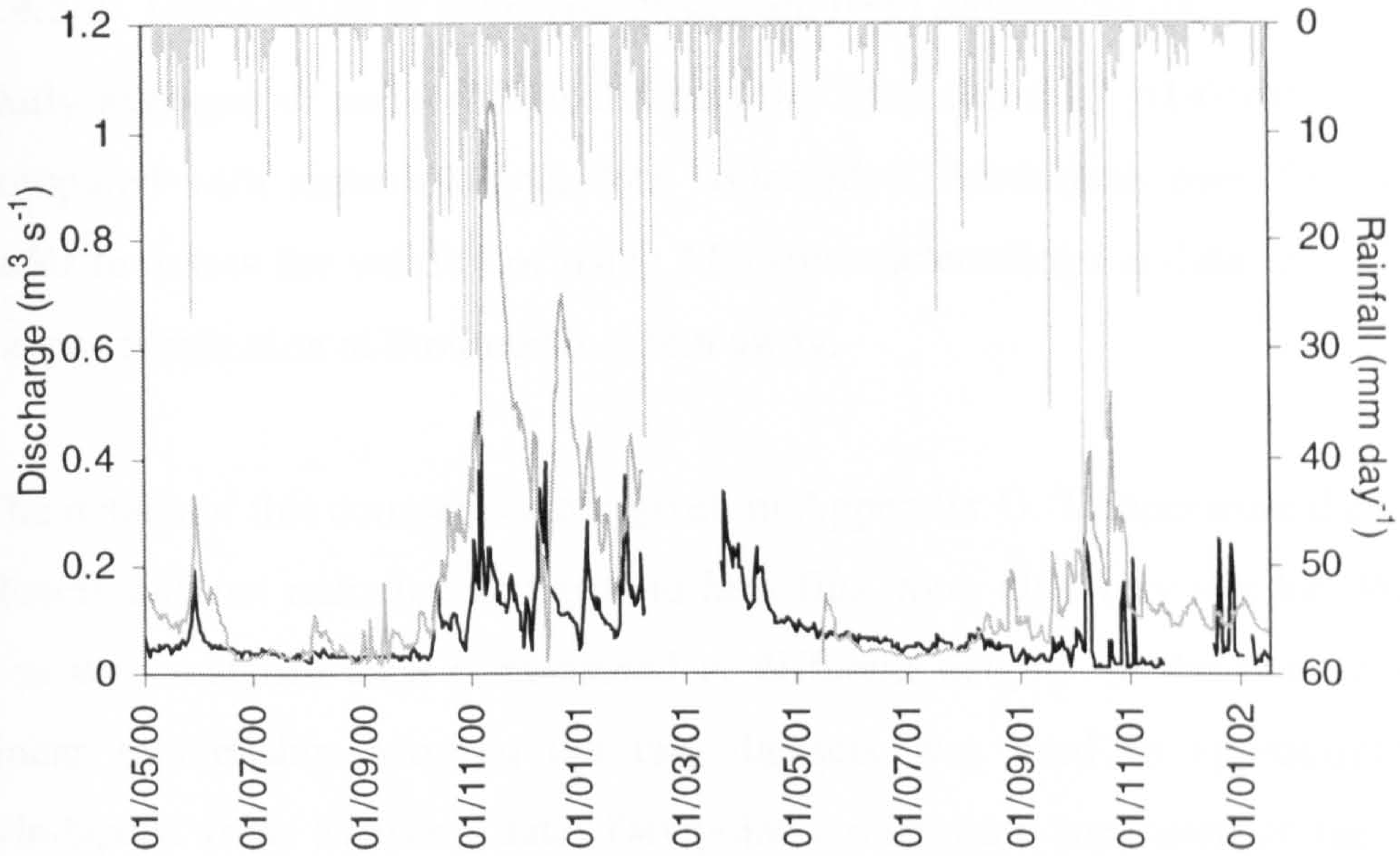


Figure 5.08 – Daily discharge at the inflow (black line) and outflow (grey line) of Stodmarsh National Nature Reserve together with daily rainfall totals (grey bars).

5.4.3 Evapotranspiration

5.4.3.1 Reference evapotranspiration

The reference evapotranspiration calculated from meteorological data from Manston can be seen in Figure 5.09.

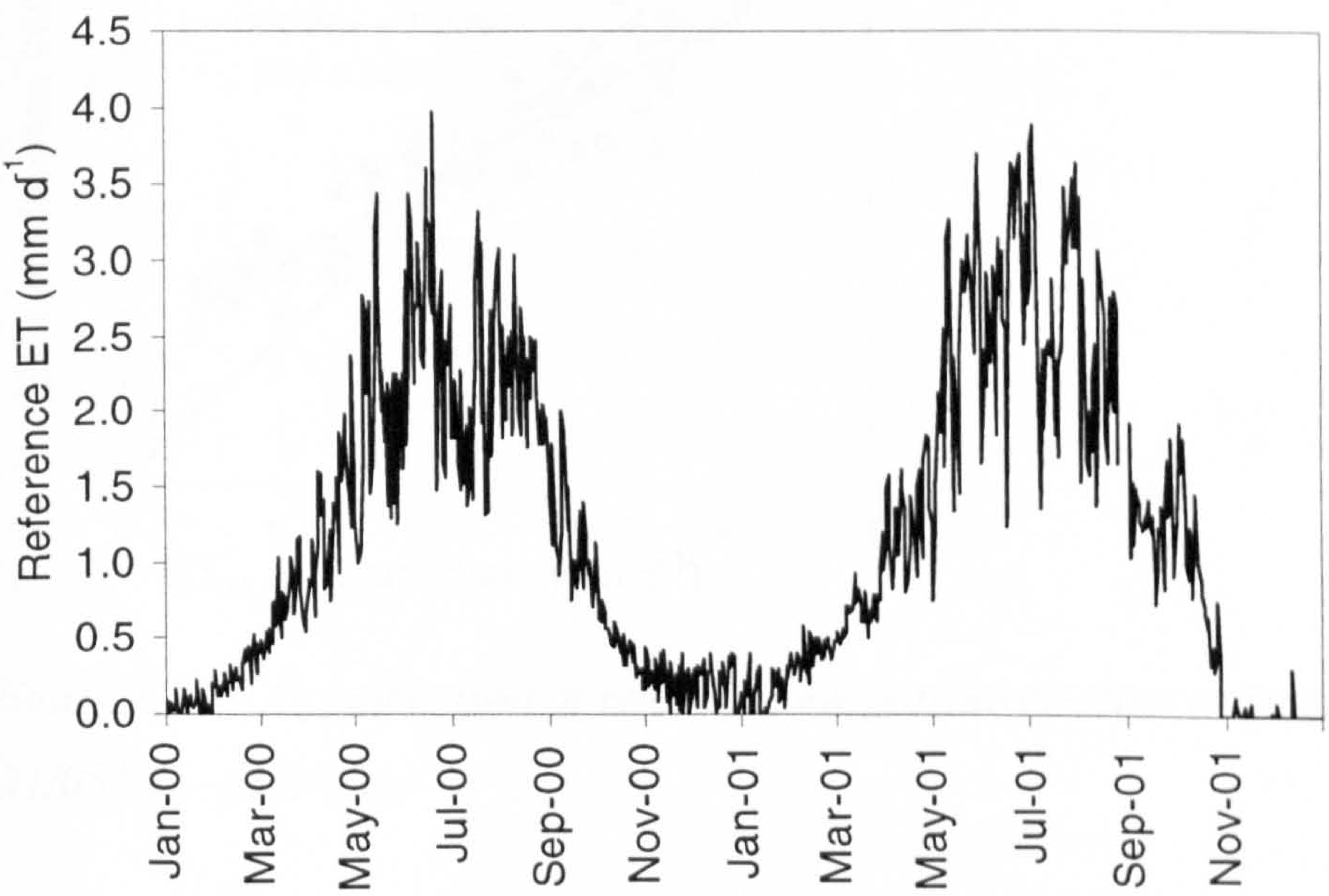


Figure 5.09 – Reference evapotranspiration 2000-2001 using data from the Manston meteorological station.

5.4.3.2 Comparison of meteorological data from Stodmarsh and Manston

Daily averages of meteorological data from Manston from 31/05/01 to 28/08/01 were compared with meteorological data collected at Stodmarsh over the same period in order to assess the validity of using Manston meteorological data to create a model of evapotranspiration at Stodmarsh, 10km away.

The details of this comparison are given in Appendix G. Temperature data matched very closely and net radiation and ground heat flux were also very similar. Windspeed was less well matched as it is measured at different heights on the two sites. However a linear relationship between the two datasets was used to approximate Stodmarsh windspeed from Manston data. Dewpoint temperature measured at the two sites was also significantly different, possibly due to the difficulties in measuring this parameter accurately and instrumental differences. Figure 5.10 summarises the comparison between reference evapotranspiration (Equation 3.14) calculated using data for Manston and reference evapotranspiration calculated using data from Stodmarsh for the same period.

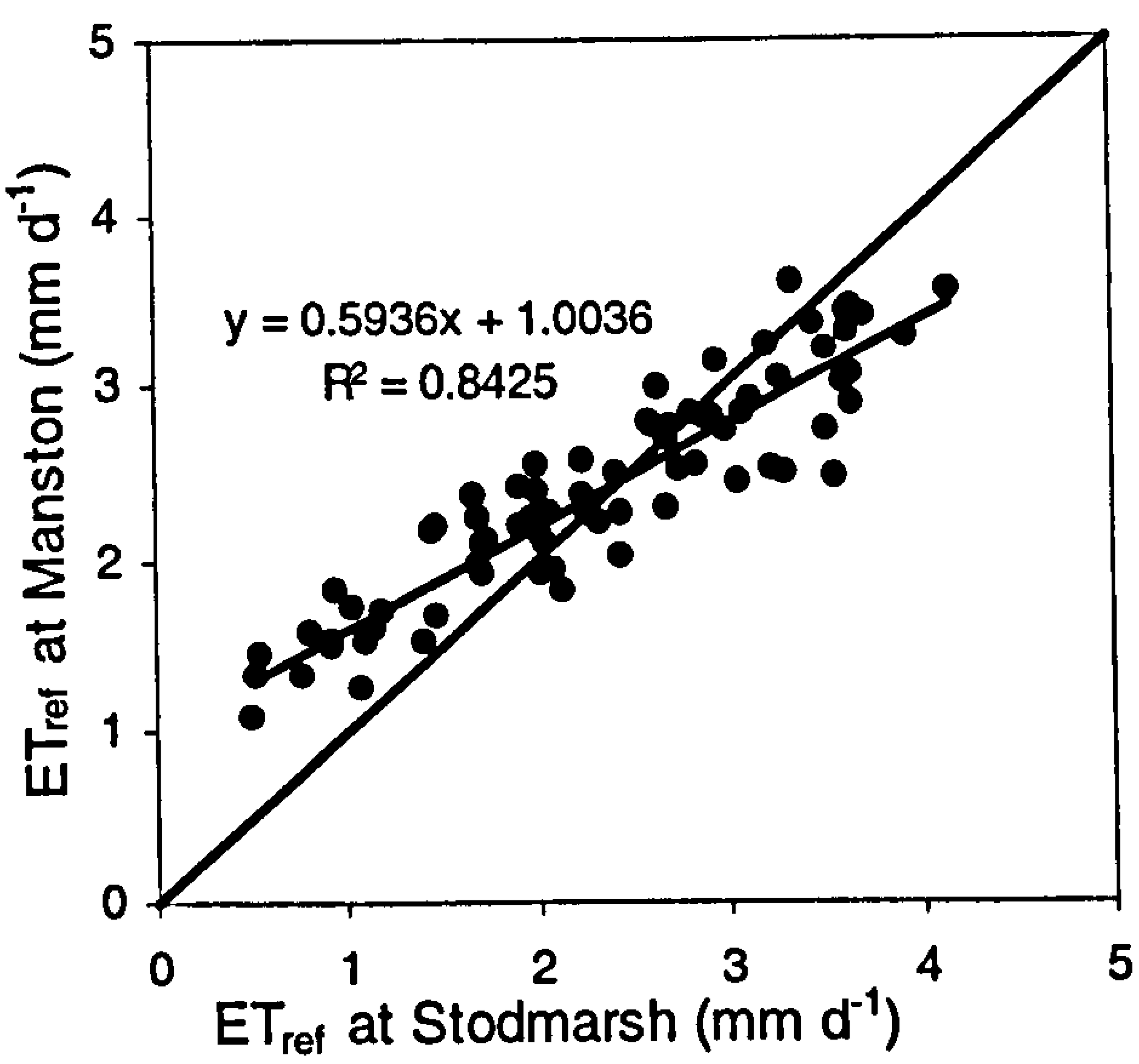


Figure 5.10 – A comparison of reference evapotranspiration at Stodmarsh and Manston (31/05/01 – 28/08/01)

The different data sets result in some scatter and the slope of the regression line of the data is significantly different from the 1:1 line. It appears that Manston predicts higher evapotranspiration than Stodmarsh when evapotranspiration is low and vice versa when evapotranspiration is high. This may be related to the differences in meteorological measurements at the two sites, mentioned above. This, however, results in very similar average evapotranspiration over the whole data set at the two sites (2.34 mm d^{-1} compared with 2.39 mm d^{-1}).

5.4.4 Change in storage

The map in Figure 5.11 shows the levelling survey and an example of water balance measurements from August 2000.

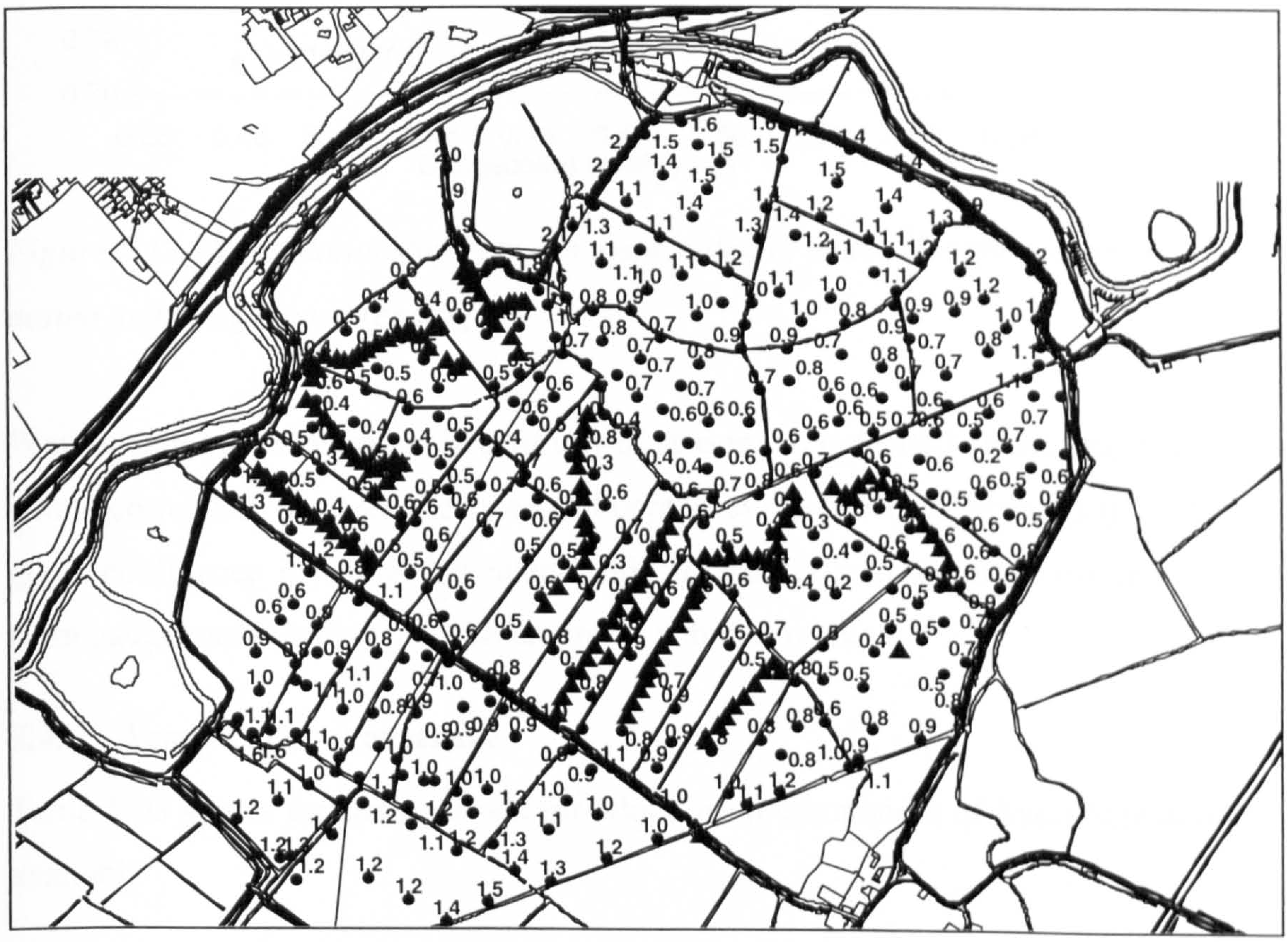


Figure 5.11 – The distribution of levelling points on the digitised map of Stodmarsh. Measurements marked with circles are levelling points, and those marked with triangles are measured depths in August 2000.

There was generally good agreement between the values of water elevation above datum calculated from the nearby levelling measurements, and those calculated using change from previous measurements.

Figure 5.12 shows the relationship between the mean value of water elevation above Ordnance datum, against the gaugeboard reading recorded on that day.

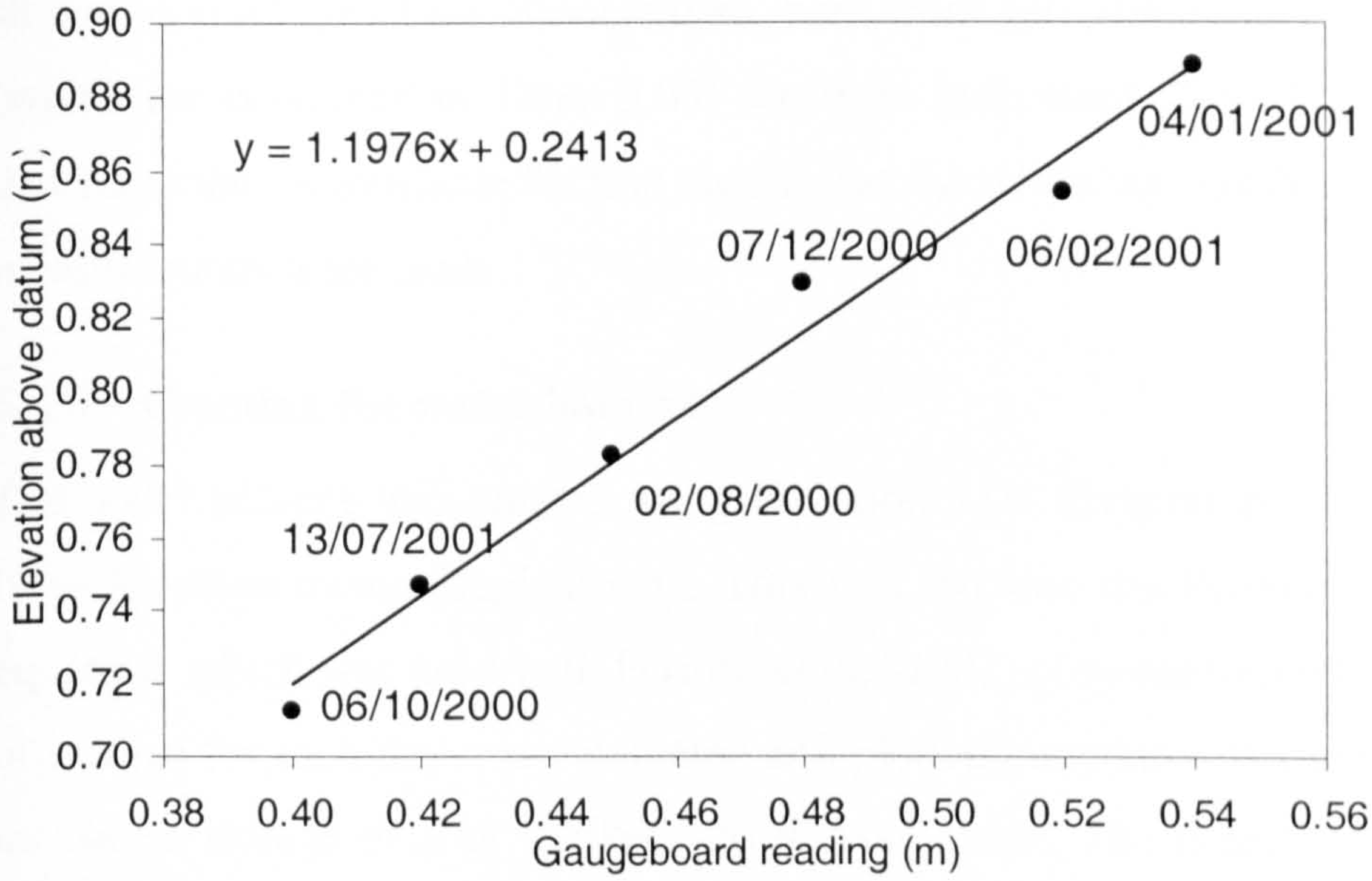


Figure 5.12 - The relationship between mean value of water elevation above Ordnance datum and gaugeboard reading

Water depths were measured in a variety of seasons as indicated on the graph. There is a strong correlation between the measured depths and gaugeboard readings (r^2 0.98). This gives confidence in using this relationship to predict average elevation above datum from gaugeboard readings, which are much simpler to take.

5.4.5 Temporal data coverage

Table 5.03 shows the time periods for which each component of hydrological data are available

Table 5.03 – Months for which hydrological data are available (shaded areas).

	2000									2001											
	A	M	J	J	A	S	O	N	D	J	F	M	A	M	J	J	A	S	O	N	D
Rainfall																					
Stream inflow																					
Stream outflow																					
Etref																					
Bowen ratio																					
Storage	O1		O4	O1	O2		O4	O1	O1	O1	O1			O 2	O 3	O4	O 3				

The first five data sources are measured as continuous daily data and are therefore shown as solid grey bars. Storage data, both from gaugeboard and direct measurements (which are combined in Table 5.03) are from individual days. A grey circle indicates that some data is available for that month and the following number specifies how many measurements were made.

5.4.6 Creating the water balance

The water balance was created using Equation 5.05. Evapotranspiration was estimated from Manston meteorological data. This was fed into the Penman Monteith reference equation, which was used with Fermor *et al.* (2001) crop coefficients. Estimated change in storage for each day was calculated and plotted cumulatively through time, using the measured storage data of 02/08/01 as a fixed point. The predicted storage volume is compared to the measured storage volumes, shown with their error bars, in Figure 5.13. There is good agreement between storage predicted from the water balance equation and that measured on site.

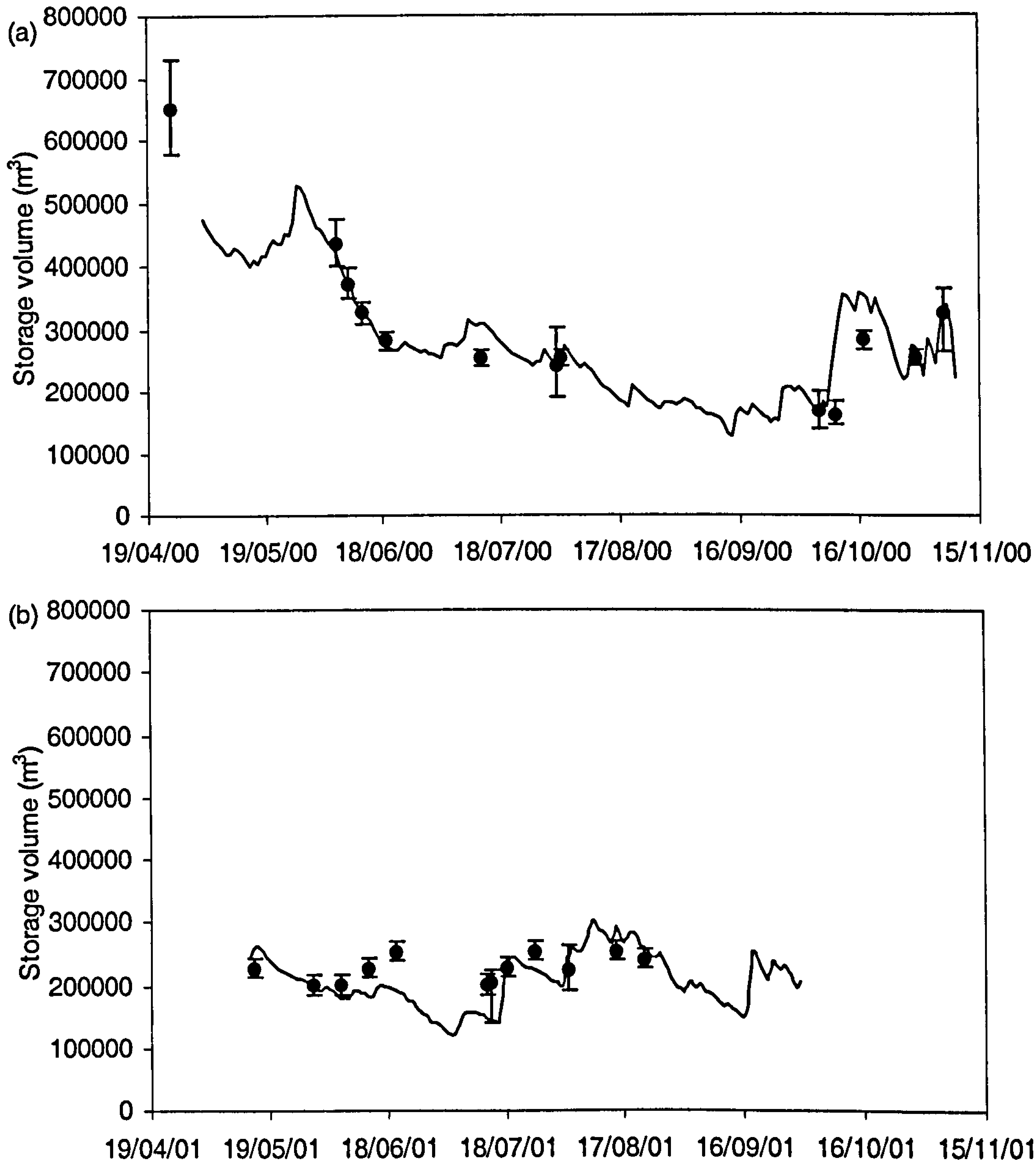


Figure 5.13 – Storage estimated from the water balance (line) and storage from measured depths and gaugeboard readings (points with 95% confidence bars) for (a) 2000 and (b) 2001.

The water balance was also used to estimate crop coefficients using Equation 5.07. This was done by working out the water balance using reference evapotranspiration then adjusting the monthly crop coefficients in order to minimise the difference between predicted change in storage and measured change in storage. This could only be done between periods for which there are estimates of storage volume data. The results of this can be seen in Table 5.04.

Table 5.04 – Kcs calculated from the water balance

Year	Period	Rain (mm)	Stream inflow (mm)	Stream outflow (mm)	Storage change (mm)	ET (mm)	ET_{ref} (mm)	*Kc*
2000	07/06-19/06	1.1	25.5	50.0	-63.7	40.2	32.2	0.80
2000	19/06-01/08	70.4	57.4	53.1	-16.2	90.9	75.4	0.83
2000	01/08-29/08	50.4	35.5	54.0	-25.6	57.5	32.8	0.57
2001	30/05-20/06	28.0	49.3	35.8	-4.4	45.6	33.3	0.61
2001	20/06-18/07	57.5	51.2	33.4	-10.6	33.4	26.7	0.80
2001	18/07-22/08	97.8	70.5	75.3	+5.3	87.7	65.8	0.75

These results are summarised in the pie charts in Figure 5.14.

Key to pie charts:

- Stream inflow
- Rainfall
- Change in storage
- Evapotranspiration
- Stream outflow
- Unknown / error

If the storage change component is on the outputs side, this indicates an increase in storage. When storage change is an input, this indicates a decrease in storage levels over the period shown.

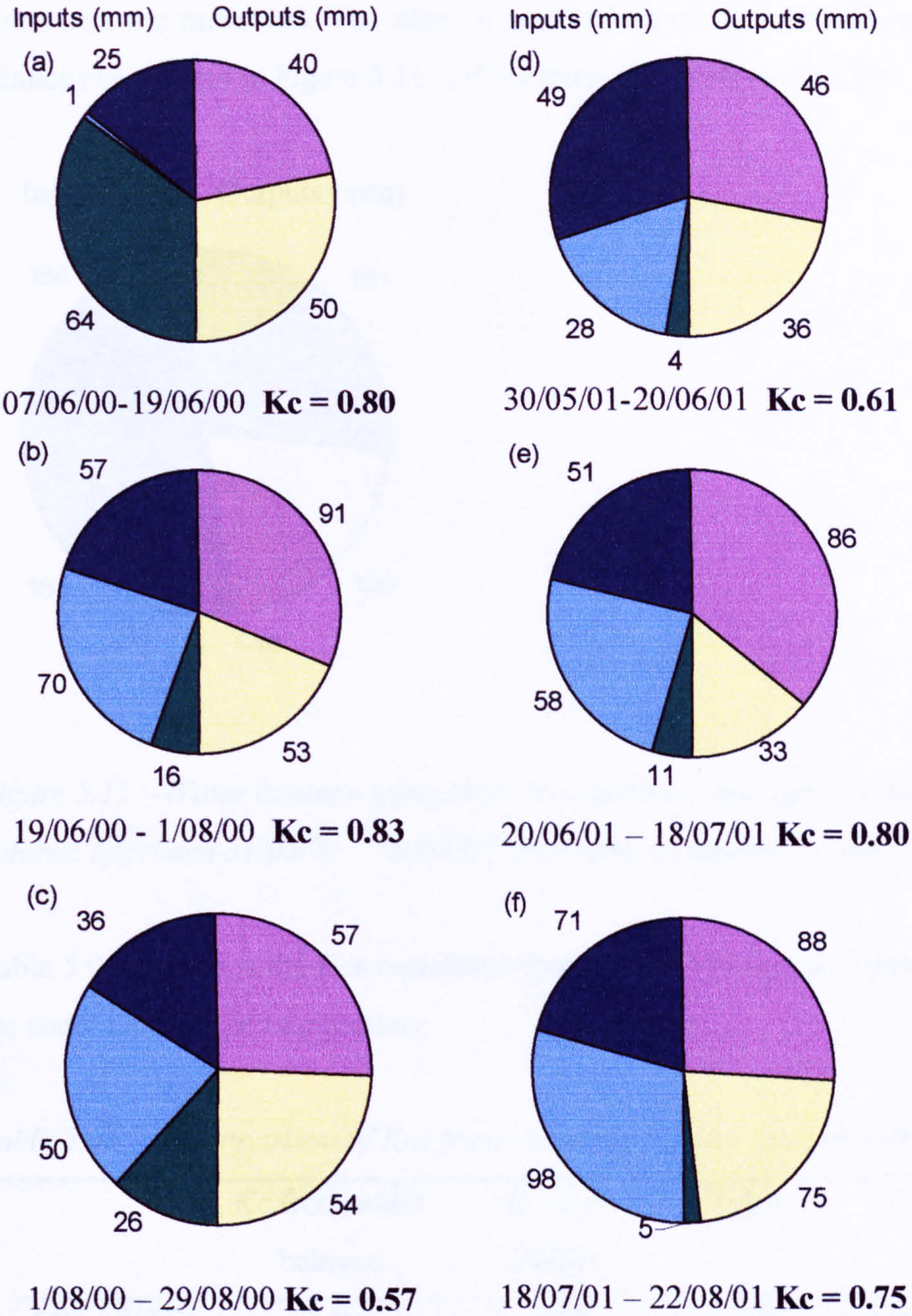


Figure 5.14 – Proportional contributions of the components of the water balance in summer 2000 (a-c) and 2001 (d-f). The figures refer to total water depth equivalent (mm).

In 2001 the water balance was also calculated using measured Bowen ratio data (see Chapter 3) for the evapotranspiration input. This removes the uncertainty associated with crop coefficient values and the differences in climate between Manston and Stodmarsh and therefore there should be a balance between inflows and outflows as all parameters are measured. The relative proportions of the components making up this balance can be seen in Figure 5.15. Inflow exceeded outflow by 1.2%.

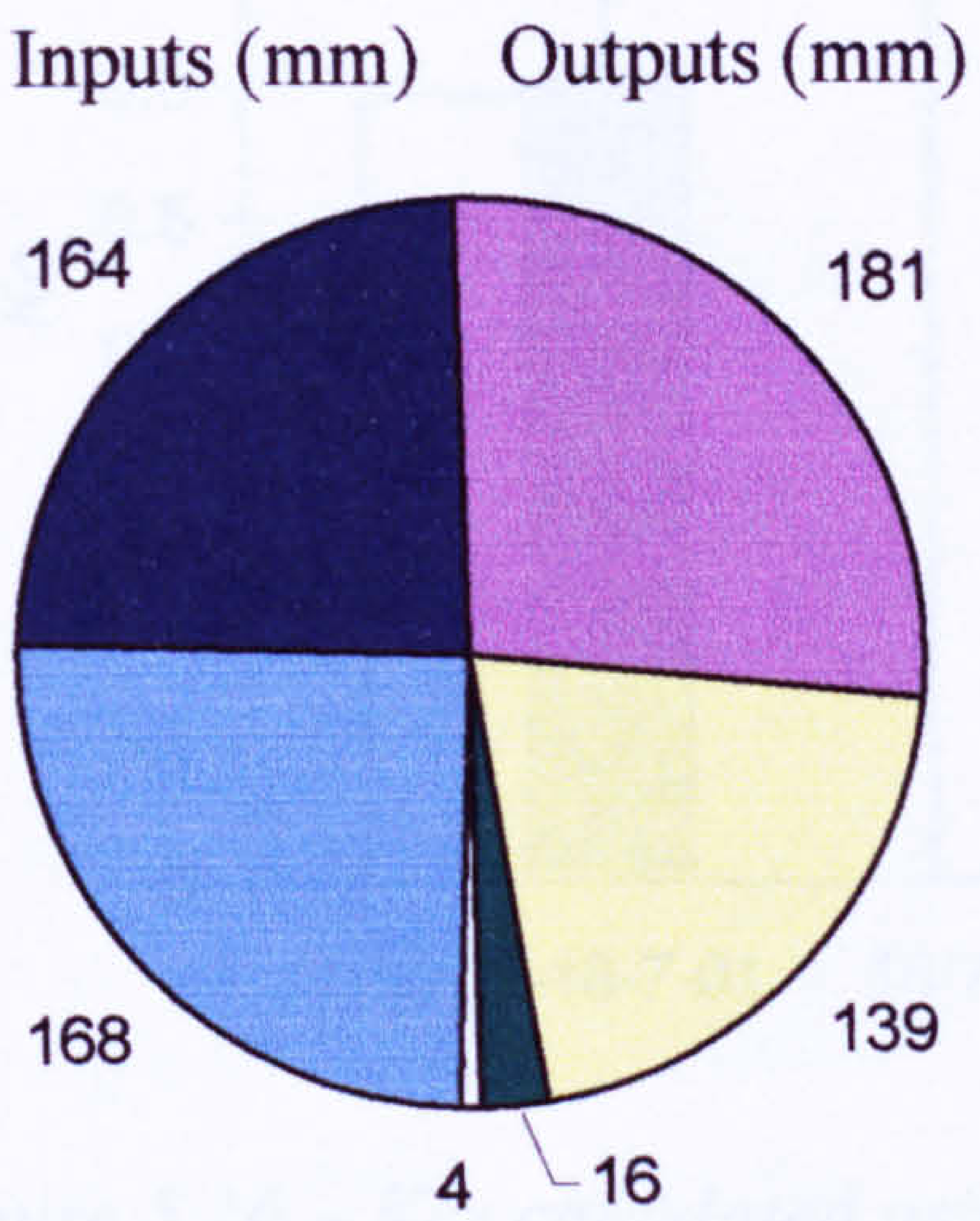


Figure 5.15 – Water balance using evapotranspiration measured using the Bowen ratio balance approach 31/05/01 – 28/08/01, including an unknown component (4 mm)

Table 5.05 compares the *Kc*s calculated from the BREB data and those calculated from the water balance for two periods:

Table 5.05 – A comparison of Kcs from the water balance and from BREB

	Kc from water balance	*Kc* from BREB	St. dev.	SE	±95%
31/5/01-18/7/01	0.59	0.60	0.301	0.043	0.087
18/7/01-22/8/01	0.75	0.70	0.331	0.055	0.111

This is expressed in Figure 5.16.

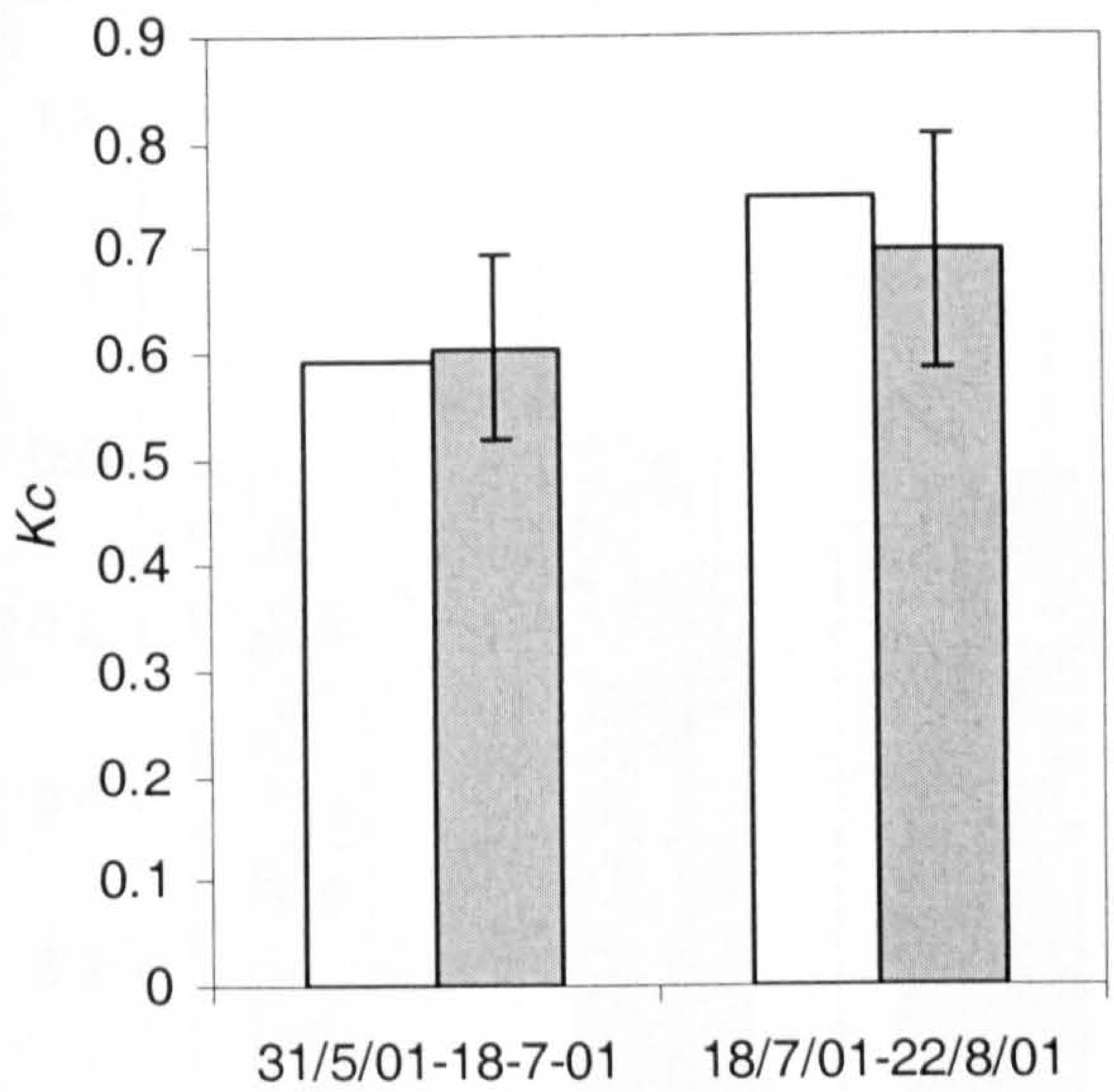


Figure 5.16 – Kcs calculated using Bowen ratio approach (grey bars with 95% confidence bars) and the water balance (white bars).

The error bars represent 95% confidence intervals. The *Kc* as calculated by the Bowen ratio approach is quite variable from day to day, hence the large confidence interval bars.

In summary, three sets of crop coefficients for the summer months have been considered: from Fermor *et al.* (2001), from back calculating the water balance and from the BREB. Table 5.06 compares these *Kc*s.

Table 5.06 – Comparison of monthly crop coefficients as calculated by Fermor et al., the water balance and the Bowen ratio energy balance method.

	Fermor *et al.*	Water balance			Bowen ratio
	Average (1994-1998)	2000	2001	Average	2001
June	0.77	0.80	0.63	0.72	0.73
July	0.86	0.83	0.80	0.82	0.68
August	0.72	0.57	0.75	0.65	0.78

These are compared in Figure 5.17.

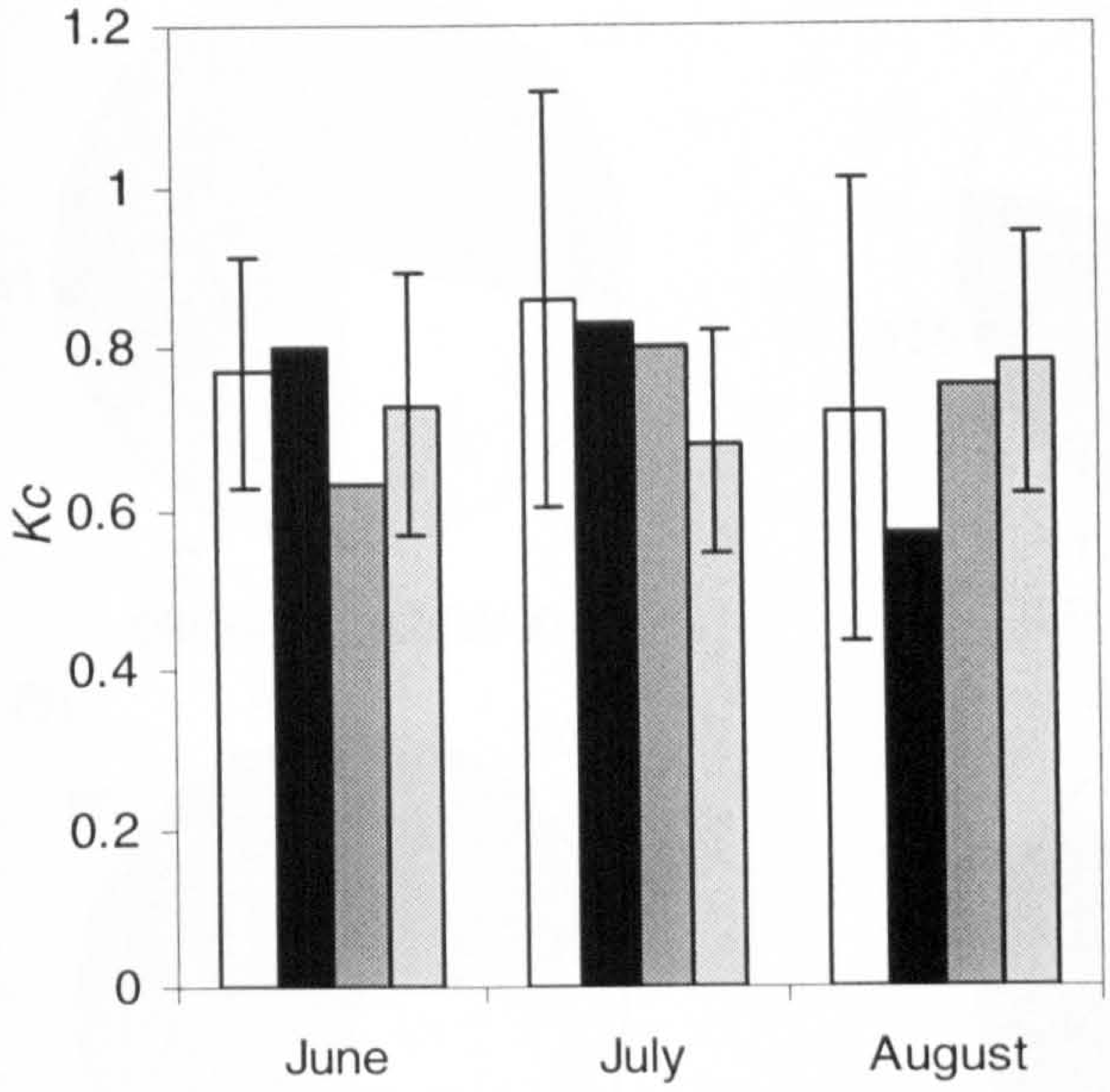


Figure 5.17 – A comparison between the average Kc of Fermor et al. (2001) (calculated 1994-1998) (white bars with 95% confidence intervals), the Kcs calculated from the water balance in 2000 (black bars) and 2001 (dark grey bars) and Kcs calculated from the Bowen ratio approach in 2001 (light grey bars with 95% confidence intervals).

It can be seen that in most cases there is no significant difference between the *Kc* values. The large error bars are due to the wide inter-annual variation found by Fermor et al. (2001) and the daily variation when using the Bowen ratio approach.

The importance of the components of the water balance varies between seasons. This can be seen in Figure 5.18 for summer 2000 to winter 2000/1. Fermor crop coefficients are used in the summer and a set coefficient of 0.5 is used in the winter.

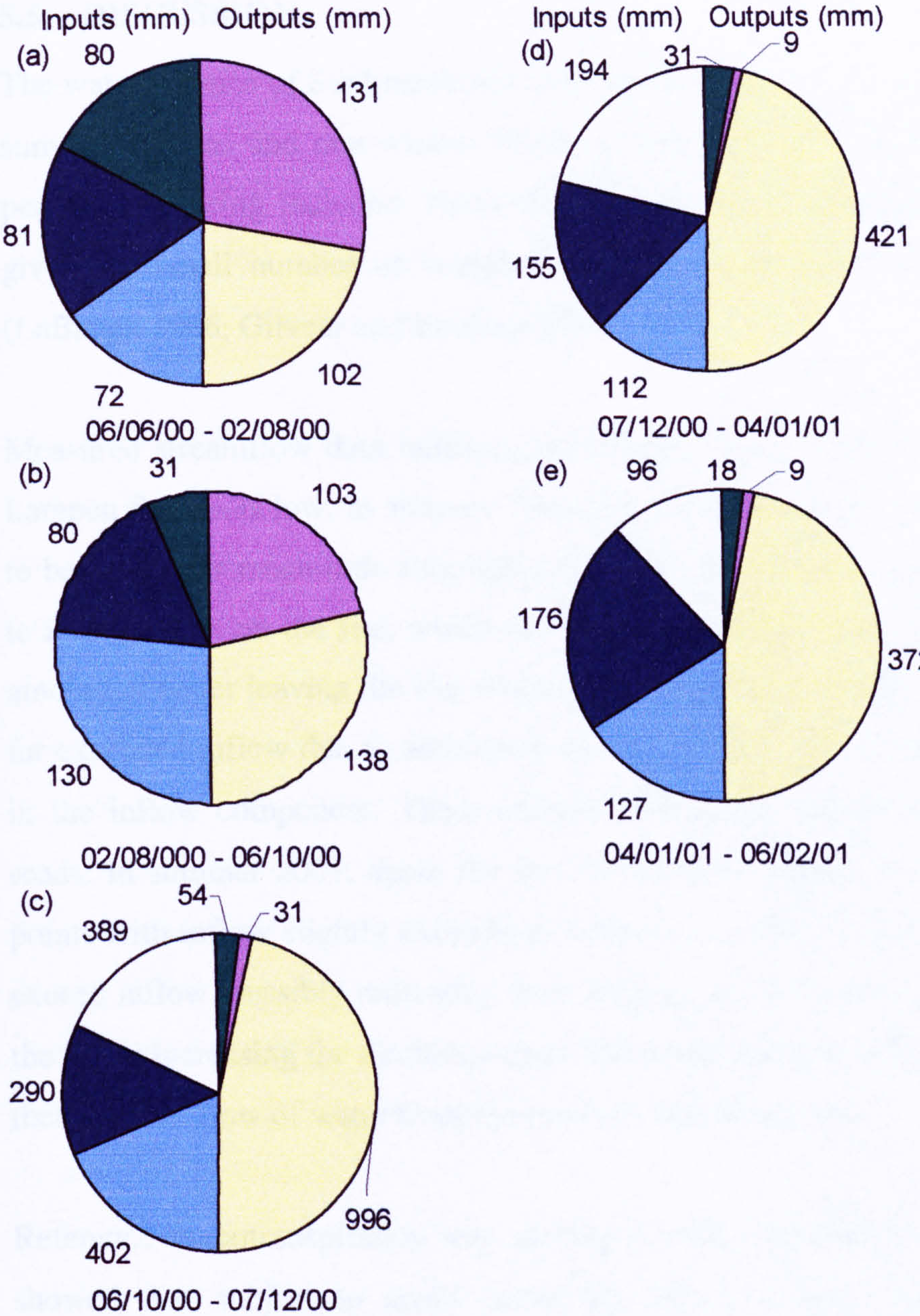


Figure 5.18 – The proportional importance of components of the water balance during different seasons of the year from June 2000 to February 2001. The figures refer to total water (mm).

5.5 DISCUSSION

The water balance of Stodmarsh National Nature Reserve has been monitored over two summer seasons and one winter. Flooding and instrument failure prevented a longer period from being included. However this remains an unusual and valuable data set given the small number of complete wetland water balances calculated in the UK (LaBaugh 1986; Gilvear and Bradley 2000)

Measured streamflow data indicates that during the summer months, discharge in the Lampen Stream is low. In summer 2000 the inflow and outflow low flow levels appear to be of similar magnitude although the outflow discharge is peakier. This may be due to management on the site, which can involve opening sluices in order to increase the amount of water leaving the site when water levels are too high. In winter 2000 outflow far exceeded inflow due to additional sources of water which were not being measured in the inflow component. These include river water and floodwater coming in over roads. In summer 2001, again the flow levels were similar at the inflow and outflow points with inflow slightly exceeding outflow. As winter approaches outflow begins to exceed inflow, possibly indicating river seepage, or alternatively simply that rainfall on the site is increasing the discharge more than evapotranspiration is reducing flow so that there is a net flow of water from the reserve into the stream.

Reference evapotranspiration was calculated using data from Manston airport. Analysis showed that subject to some correction, this is valid. Although some individual variables show quite significant differences between measurements at Manston and Stodmarsh, particularly windspeed and vapour pressure, the overall evapotranspiration measurements show an acceptable level of agreement. There is a likelihood of slight underestimation at times when evapotranspiration is high and overestimation when it is low on a daily basis but these should balance out over longer periods of time.

In both 2000 and 2001, the water balance model (Equation 5.03) appears to be successful at predicting the change in storage. It is necessary to be cautious about these predicted changes in storage, as they are essentially residuals of the water balance and therefore contain all the errors of the other data inputs. However there is a good match between the storage predicted using the balance and that measured directly or calculated from gaugeboards. This gives confidence that all the necessary components of the balance during these periods were measured and that the error levels were acceptable. This confidence is increased by the closeness of the balance between inputs and outputs when all five components of the water balance are measured, which results in the input volume equalling the output to within 1.2% (Figure 5.15).

The years investigated had higher than average rainfall. However this did not apply uniformly to every month. April 2000 was an extremely wet month followed by a very dry May leading to rapid dropping of water levels. June to September were closer to average whilst October and November were extremely wet signifying the onset of large scale flooding. Wet weather continued throughout the winter culminating in February 2001 with the worst floods seen in over 50 years when water was 2 m above the surface at some points on the reserve. In April the rainfall returned to more usual levels with June being dry and August and September wetter again. October was very dry. The precipitation patterns can be seen reflected in the change in storage graphs. The high rainfall levels in April 2000 led to high levels of water on the reserve. However the higher temperatures (and therefore higher evapotranspiration rates) and dramatically lower rainfall levels in May and June led to a rapid fall off of water from the end of May which continued for the rest of the summer as evapotranspiration rates exceeded rainfall, until rainfall started to increase again. An estimate of change in storage could not be created after November as water started to breech roads and the river bank meaning that water was moving outside our measurement points.

The change in storage was much smaller in 2001 than in 2000. This can be explained by the fact that despite very heavy rainfalls in February 2001, the reserve recovered quickly from flooding and water levels subsided during spring time, under closer to average rainfall conditions, to a much lower level in May 2001 than it had been a year previously. Water levels remained fairly low, increasing in late July and August with the wetter weather.

The pie charts in Figure 5.14 give further clues as to the reasons for the differences in the patterns of water storage between 2000 and 2001. Comparing Figures 5.14 a and d which show the water balance in June of each year, it can be seen there are very large differences in the importance of the different components of the water balance. Absolute totals must be weighted by the lengths of the periods compared (necessarily different due to limited storage measurements) but it can be clearly seen that in 2000 rainfall was very low during this period, leading to low streamflow and a very rapid reduction in storage levels on the site. Evapotranspiration was also higher in 2000 contributing to the reduction in storage levels. However mid June to the end of August was much more similar between the two years in terms of rainfall quantity and the response of storage levels. Divergence occurs again in August (Figures 5.14 c and f) with the storage levels still dropping under lower rainfall and significantly higher evapotranspiration levels in 2000 (2.0 mm a day as compared to 1.7 mm day^{-1}).

Although the summer period is most important in terms of water deficits, it is also instructive, and indeed necessary in terms of modelling, to study the winter water balance. Figure 5.18 shows the progression of the balance throughout the winter months. During the winter there is an unknown component included in the balance as the input and output are not equal – the output is much greater than the input. Until the 6th October 2000 storage decreases and evapotranspiration is still a significant part of the water balance (Figure 5.18b). However after this (Figure 5.18c) there is a big change. Evapotranspiration becomes very small and storage starts to increase. This pattern continues throughout the rest of the winter until February. During this period the input of water onto the site is fairly evenly divided between rainfall, inflow from the Lampen Stream and an unknown component. On-site observations indicate that this component is likely to be river water from the Great Stour, which, due to the unusually high rainfall from October 2000 onwards resulted in high river discharges, initially causing seepage of water through the bund and in February the over-topping of the bund. It is difficult to say whether this river component is an entirely flood-related phenomenon or whether some input could be expected in all years. Conversations with site managers have indicated that seepage is a common feature of the site when river levels are high. Greater outflow than inflow at the start of winter 2001 indicates that there may also be river seepage at this point.

During the period when all components of the water balance were measured, Stodmarsh NNR had fairly equal contributions of surface water and precipitation and negligible groundwater input. Using the classification triangles of Brinson (1993a,b) and Lent *et al.* (1997) Stodmarsh can be classified as to its wetland "type". According to Brinson (1993a) Stodmarsh is a bog. This is slightly surprising as Keddy (2000 p.18) defines a bog as "a wetland community dominated by *Sphagnum* moss, sedges, Ericaceous shrubs or evergreen trees rooted in deep peat", not a description of Stodmarsh. Stodmarsh fits better the definition of a marsh: "A wetland community that is dominated by herbaceous plants that are usually emergent through water and rooted in hydric soils but not in peat. Examples include ... reed (*Phragmites*) beds" (Keddy 2000 p.18). The triangles of Lent *et al.* (1997) have more possible types and do not attempt to attach labels to the classifications. Here Stodmarsh is identified as class IV for both inflow and outflow, being dominated by two out of three sources.

The overall picture seems to indicate that it is possible to model the change in storage on Stodmarsh National Nature Reserve using a simple water balance model, at least during the summer months when river water does not appear to feature in the hydrology of the site. In winter the creation of models is more problematic due to the fact that the threshold river discharge for input into the reserve to occur (if indeed there is one at all) is unknown. The research also indicates that there is enormous variety of possible hydrological regimes that could occur on the site. Here, two summers that would both be described as "wet" are seen to result in very different water level regimes.

All balance studies must be viewed with a note of caution as corresponding errors on opposite sides of the balance may cancel one another out making the balance appear to give more confidence in the output data than it really should. An imbalance (or even a balance) can be due to both errors in measurement and undetected processes (Gilman 1992). There are potential errors in every data input to the model. More attention is given to the problem of errors in the balance and subsequent modelling in Chapter 8.

There is also the possibility of significant fluxes occurring on the site that have not been considered. The issue of seepage from the river is an example of this. Other processes that have been considered negligible in the assumptions that could possibly have more importance than has been accredited to them are groundwater and diffuse overland flow.

The contribution of a simple water balance to our understanding of wetland hydrology is that it expresses the relative importance of various water sources and hydrological processes. However although a water balance is an important first step in understanding the hydrology of a wetland site, in order to be able to carry out detailed management and analysis more than a simple budget is required. For a better understanding it is necessary to look deeper into the workings of the system and understand the interactions of the wetland with rivers, the atmosphere and groundwater and only when the detail is filled out can it become a sound basis for decision making (Gilman 1994).

In light of the emphasis on the measurement of evapotranspiration in this research, the water balance has created interesting results. In the first instance the crop coefficients of Fermor *et al.* (2001) from Walton Lake were used (in winter a constant coefficient of 0.5 was chosen as the Fermor coefficients varied a lot, for example 1.09 in January and 0.46 in February). In summer 2000 these fitted the water balance well indicating that there may be comparability between the two sets of results. In 2001 the fit with the Fermor coefficients was less good but still acceptable. The crop coefficients created using the Bowen ratio method are similar to those of Fermor especially in June and August, although there is a significant difference in July. However when the measured Bowen ratio data is used in the water balance, inputs and outputs are almost equal as demonstrated by Figure 5.15.

The water balance can be used as an additional source of crop coefficients, by setting evapotranspiration instead of change in storage as the residual component. These *Kc*s are produced over periods that depend on when storage data was available. Again, caution must be exercised, as these are residuals. It can be seen that in general these compare well both with Fermor and with the *Kc*s from the Bowen ratio approach, though there are discrepancies, as there are discrepancies from year to year, for example in August 2000 the *Kc* is 0.57 and in 2001 it is 0.75. This indicates potential difficulties with modelling, as using *Kc*s predictively relies on a single crop coefficient that can be used from year to year for a particular period of the year. However from this we have confidence that we are in the correct range for crop coefficients and that crop coefficients are smaller in summer months than the figure of 1.4 that has previously been suggested (Hawke and Jose 1996; Bardsley *et al.* 2001).

The water balance also shows that as evapotranspiration is such a small part of the water balance from October and throughout the winter, the actual value of crop coefficients will be much less important than in the summer. The fact that evapotranspiration has not been measured in winter will therefore not be a great problem when it comes to modelling.

5.6 CONCLUSIONS

The water balance has been measured over two seasons. The data for each component of the water balance has been deemed acceptable. The use of a simple water balance equation resulted in estimations of the change in storage for summers of 2001 and 2002 that matched well with measured and calculated water storage data. When evapotranspiration measured using the Bowen ratio technique was included (meaning that all five components of the water balance were being measured and there was no residual component) the water balance was closed to within 1.2%. The water balance was also rearranged so that crop coefficients could be found from it and these agreed well with those calculated from the Bowen ratio technique and also with those estimated at Walton Lake by Fermor *et al.* (2001). The measured water balance of Stodmarsh National Nature Reserve appears to be a reasonable basis on which to construct a model to predict future change in storage.

CHAPTER 6 – Modelling the Hydrology of Stodmarsh National Nature Reserve

6.1 INTRODUCTION

A water balance has been created using the measured hydrological data from Stodmarsh National Nature Reserve (NNR) and a simple model was found to be successful (Chapter 5). The next step is to model the components of the water balance in order to predict future fluxes through and storage within the wetland under different meteorological conditions. There are an infinite number of possible future climate scenarios. Therefore, historical data are being used to determine probabilities of the recurrence of extremes of key meteorological variables and from these the recurrence of water levels below those considered optimum for the conservation objectives of the site.

6.1.1 Water regime management

Water level management is the single most important factor in conserving a reedbed for wildlife (Ward 1991). It is not sufficient simply to maintain a constant water level throughout the year, but a regime must be achieved that is in keeping with the wildlife objectives of the site and the amount of water available. Water level variation contributes to habitat diversity, flushes the system of toxins and decreases the rate of sedimentation (Gilman 1998). There are a number of factors that must be held in balance when determining optimal water levels. Some bird species require shallow water for feeding and nesting, whilst it is important to maintain a substantial depth of water throughout most of the year so that the *Phragmites* bed is not invaded by other species. The focus of reedbed management at Stodmarsh NNR is to encourage Bitterns to breed. Bitterns require a reed-water interface for feeding and wet reedbed 100-300 mm deep for breeding and feeding (Andrews and Ward 1991). The Bittern's leg length confines them to water less than around 250 mm deep. They nest in shallow water typically around 100 mm deep amongst reeds. However, there are also other important species on the site. Summer flooding provides Bittern feeding habitat and inhibits scrub invasion but may prevent the development of the carr-reed interface that

Cetti's Warbler requires and remove nesting areas for Bearded Tits which require dry areas.

A variety of water level regimes can be implemented on a reedbed. Water may be maintained at high levels all year round, or dropped after the bird breeding season in June and raised again in autumn, or it may be kept low in autumn and kept dry over winter and be re-flooded in the spring. Management practices may also have an influence, for instance levels may be drawn down in autumn to allow reed cutting. Once a regime is decided upon, the ability to replicate it from year to year is desirable as sudden changes in water level can be damaging (Haslam 1970).

A similar regime to that recommended by Hawke and Jose (1996) is used at Stodmarsh. Water levels are raised on site in spring to a maximum surface depth of 200 – 300 mm (around late March to early April). The surface water is maintained in the range of 100 – 300 mm throughout the summer. The water drops down over the summer as evapotranspiration occurs and levels then rise again in the autumn. Hawke and Jose (1996) recommend that optimum levels for reedbeds are 50 – 300 mm in summer and less than 1 m in winter. Levels of 300 mm allow Bitterns and other wildlife to use the reedbed for winter feeding. If water levels are considered too low, additional water could be abstracted and added to the reserve around December. A typical annual hydrological profile would be sigmoidal in shape (Figure 6.01).

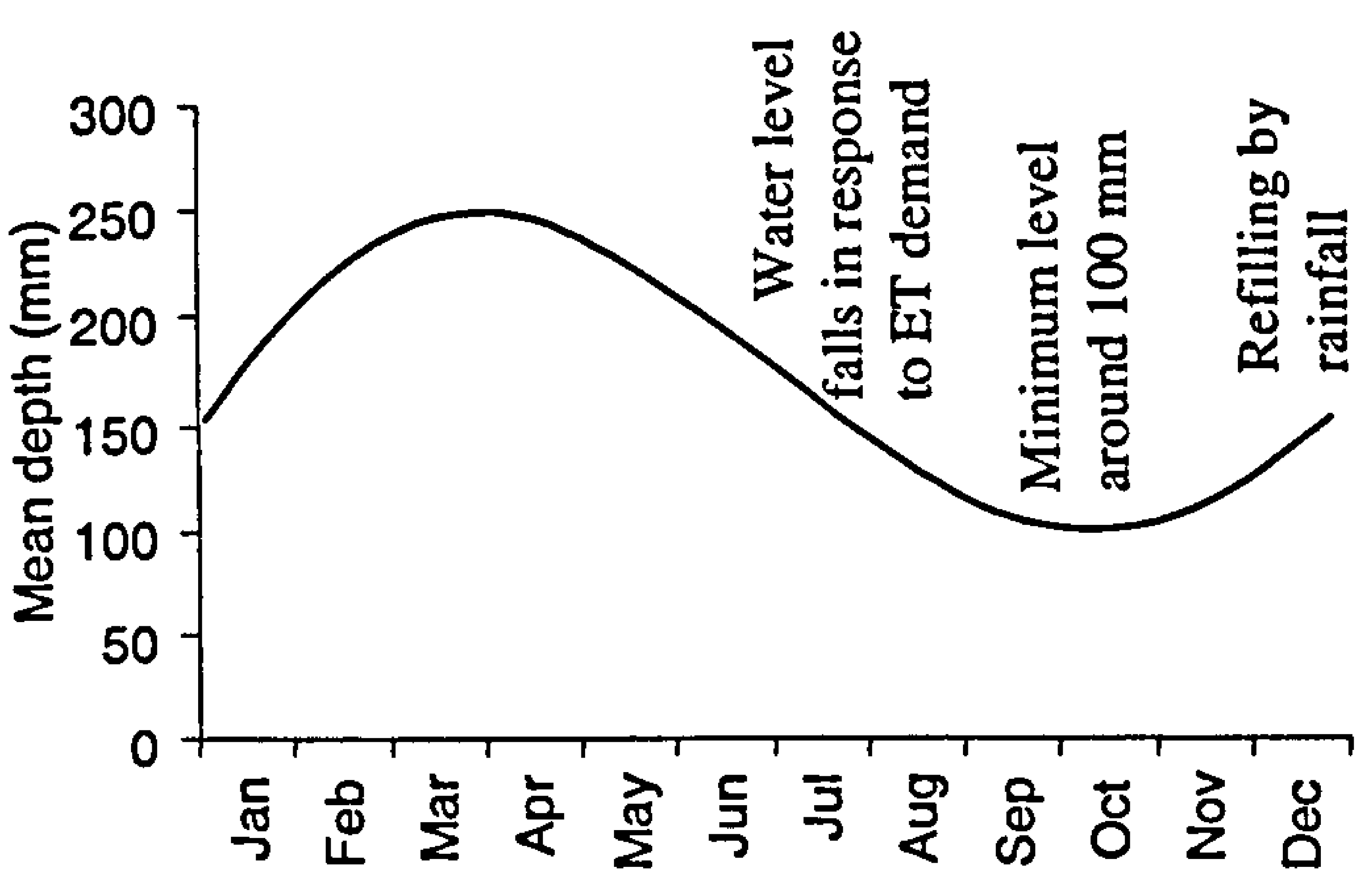


Figure 6.01 – A typical annual change in mean storage depth for Stodmarsh National Nature Reserve

6.1.2 The water balance model

The water balance model was described and verified in Chapter 5. It is:

$\Delta V/\Delta t = (P_n + S_I + R_v) - (ET + S_o)$ (after Mitsch and Gosselink 2000) (6.01)

where $\Delta V/\Delta t$ is change in volume of water storage in a wetland per unit time, P_n is precipitation, ET is evapotranspiration, S_I is surface inflow, S_o is surface outflow and R_v is river input. It was found that the change in storage predicted using this model compared well with measured storage during the summer months. However in winter there was a large unknown component (approximately of the same magnitude as precipitation or stream inflow). This was thought to be input from the River Great Stour. Evapotranspiration is a very important component of the water balance in the summer months but becomes fairly insignificant after October.

6.1.3 Use of wetland hydrological models

There have been a number of examples of the use of models to assess hydrological requirements of wetlands, both for management applications on existing wetlands and in feasibility studies of proposed wetland creation schemes. Models can be simple single rainfall event models, stochastic models giving predictions of a particular parameter or models that attempt to simulate all or the majority of the components of the water balance (Arnold *et al.* 2001).

Many of the simplest wetland hydrological models are based on water balance studies and therefore have been cited in section 5.1.3 of Chapter 5. Souch *et al.* (2000) created a long-term wetland water balance model in order to predict the feasibility of wetland creation. Thirty years of MORECS data were used to estimate mean rainfall, ET and potential soil moisture deficit and the hydrology of the catchment was gauged. A model was used to calculate water requirements of various wetland habitats and found that a reedbed in the East of England requires an average of 287 mm of water in summer and 312 mm of water in winter.

Haycock (personal communication 2001) also created a water balance model of a reedbed in preparation for the creation of a reedbed at Needingworth quarry, a gravel pit. This will involve abstracting water from the River Ouse. However, as the reedbed has not yet been created, there is no measured data against which to calibrate the model. The model used was very simple as there are no inflows or outflows other than evapotranspiration and rainfall and the area will be sealed by a clay bund. The water levels after abstraction will be 600 mm in winter and these will drop to 100 mm in summer.

Other authors have used more complex models. MODFLOW, a groundwater simulation model has been applied to several wetlands. MODFLOW is a modular three dimensional, steady state, finite difference groundwater flow model. Bradley (2002) used it to model a UK floodplain wetland and Gilvear *et al.* (1993) and Wilsnack *et al.* (2001) modelled the Everglades. They considered that MODFLOW was limited as surface flow is often simulated by assigning wetland areas very high hydraulic conductivities, so a specific wetland package for MODFLOW was developed and tested resulting in a good fit.

Su *et al.* (2000) adapted a standard semi-distributed hydrological model (SLURP) to predict the water level regime of small prairie wetlands. This model is used more commonly on large river basins which include wetlands as in the work of Kite (2001). Water balances were carried out on different areas of the site. They found that spring snow-melt and summer rainfall were the main water sources. The prediction of water levels was adequate. Walton *et al.* (1996) created the Wetlands Dynamic Water Budget model based on water balance of the Black Swamp wetlands in Arkansas, USA. It attempts to include processes found in various wetland types. The model is divided into three major modules – surface water, vertical processes and horizontal groundwater flow. It also includes channel and over bank flows, tidal forcing river inflows, flooding and drying and was calibrated successfully against measured river flow data.

Sun *et al.* (1996) modelled a coastal pine flatwood wetland using a computer simulation model called FLATWOODS, a distributed model divided into cells and requiring detailed spatial information about the catchment which is usually obtained through GIS. This was found to work with sufficient accuracy. In a subsequent study (Mansell *et al.* 2000), the same wetland was modelled using a multi-dimensional model for variably saturated water flow in porous media called WETLANDS. Evapotranspiration was estimated separately for open water, soil and transpiration. Observed and estimated water table and ponding levels were compared. Arnold *et al.* (2001) successfully used a modified Soil and Water Assessment (SWAT) model to investigate whether water sources were sufficient to maintain a proposed wetland. SWAT is a complex conceptual distributed model and data requirements are high but it was found that wetland water requirements were accurately simulated for over a decade.

6.1.4 Streamflow modelling

Stream discharge is commonly modelled using rainfall-runoff models, of which there are several types. Deterministic models attempt to simulate all the processes in the catchment and involve many complex equations. Many parameters are needed and the models are often limited by their exacting data requirements and the difficulty of relating theoretical hydrological equations to real heterogeneous catchments. Conceptual models make simplifications of reality and when they attempt to represent many sub-processes they can also become very complex with a high data demand. Parametrically parsimonious conceptual models (PPCM) where all parameters are correlated and precisely identifiable try to avoid these problems. PPCMS can perform as well as or better than complex models, have a lower data demand and are easier to use (Littlewood 2001). Internal model storage and processes become integrated so that they are no longer directly measurable properties (Liden and Harlin 2000).

6.2 OBJECTIVES

The objective of this chapter is to create a workable model of the water balance of Stodmarsh National Nature Reserve using available historical data sets as examples of potential future climate scenarios. This will be used in order to predict the change of storage on the site in future years under various meteorological conditions, and particularly under different precipitation regimes. This will enable estimates to be made of the frequency and quantity of water that will need to be abstracted and added to the site to keep it in optimal condition.

6.3 CALIBRATION AND VALIDATION DATA

6.3.1 Study site

The model is based on experimental work carried out at Stodmarsh National Nature Reserve, Kent, UK (51°19'N, 1°12'E, elevation <5 m) in 2000-2001 (see Chapter 5). The site has fairly simple hydrology with inputs from the Lampen Stream and precipitation. The site is underlain by London clay over Woolwich and Thanet beds which have a very low permeability and therefore groundwater percolation is considered insignificant. The outputs of water are stream outflow and evapotranspiration. A diagram of the site can be seen in Figure 5.04. The site is composed of around two thirds reedbed and one third wet grassland. The Lampen Stream has a small catchment, its source being a spring only 6 km from the point where it enters the reserve. It flows into the River Little Stour soon after it leaves the reserve. The two rivers in the area, the Great Stour and the Little Stour have courses that run roughly parallel to the Lampen Stream on either side of it, constricting the catchment still further. The catchment consists mainly of agricultural land – a mixture of arable land, pasture, orchards and some woodland. The topography of the area is fairly low and flat, the reserve being at less than 10 m above sea level at its lowest point and the catchment watershed is at around 40 m.

6.3.2 Rainfall data

As described in section 5.3.3 there are no raingauges positioned on the site. However within a 7 km radius of Stodmarsh are 6 currently monitored raingauges with data available (Table 5.01). Thiessen polygons were drawn using these stations (see Figure 5.02) around these raingauges and the results showed that the reserve fell entirely within the polygon for West Stourmouth raingauge (latitude 51.320° N, longitude 1.229°E, elevation 9 m). This gauge is around 1 km from Stodmarsh and has daily data from 1964. However when modelling using historical data the additional raingauges close to Stodmarsh, no longer monitored, become available. These are detailed in Table 6.01.

Table 6.01 – Details of non-current raingauges close to Stodmarsh

Station	Latitude and Longitude (°)	Elevation	Distance from Stodmarsh
Westbere	51.307°N, 1.158°E	9 m	2 km
Elmstone	51.298°N, 1.246°E	5 m	4 km

The Thiessen polygons were redrawn including the above two stations and it was found that the reserve fell partly within the Westbere polygon (Thiessen weighting 0.23), and partly within West Stourmouth (Thiessen weighting 0.77). There are data from Westbere for 1967-1986 so these weightings were used for this period. The reminder of the time, West Stourmouth data were used alone.

The West Stourmouth raingauge had some missing data. This was infilled using the equations of regression lines with data sets from other rainfall stations. Regressions were carried out between West Stourmouth and all other current raingauges and the relationship with the greatest r^2 value was used (Table 6.02). The regressions were repeated over 5 different time periods within the overall period 1964-2002 to account for data availability (only two out of the seven stations had continuous data availability, one of which being West Stourmouth) and also for the fact that relationships between stations may change over time. Table 6.02 below shows the periods used and data availability. The divisions coincide with the start or end of a station (Herne Bay was not included).

Table 6.02 – The figures are r^2 values of the regression with West Stourmouth (which has continuous data availability). The strongest correlation (bold) was used to interpose data. Grey areas indicate that no data was available.

	Garrington	Canterbury	Elmstone	Westbere	Bekesbourne	Sturry
1964-67		**0.85**	0.82			
1967-69		0.76	**0.94**	0.84		
1969-86	0.84	0.85	**0.92**	0.86		
1986-94	0.84	0.85	**0.91**			
1995-98	0.82	**0.85**				
1998-02	0.71	0.78			**0.80**	0.76

Figure 6.02 shows the annual total rainfall at Stodmarsh for 1964 –2001, based on West Stourmouth data combined with Westbere when available.

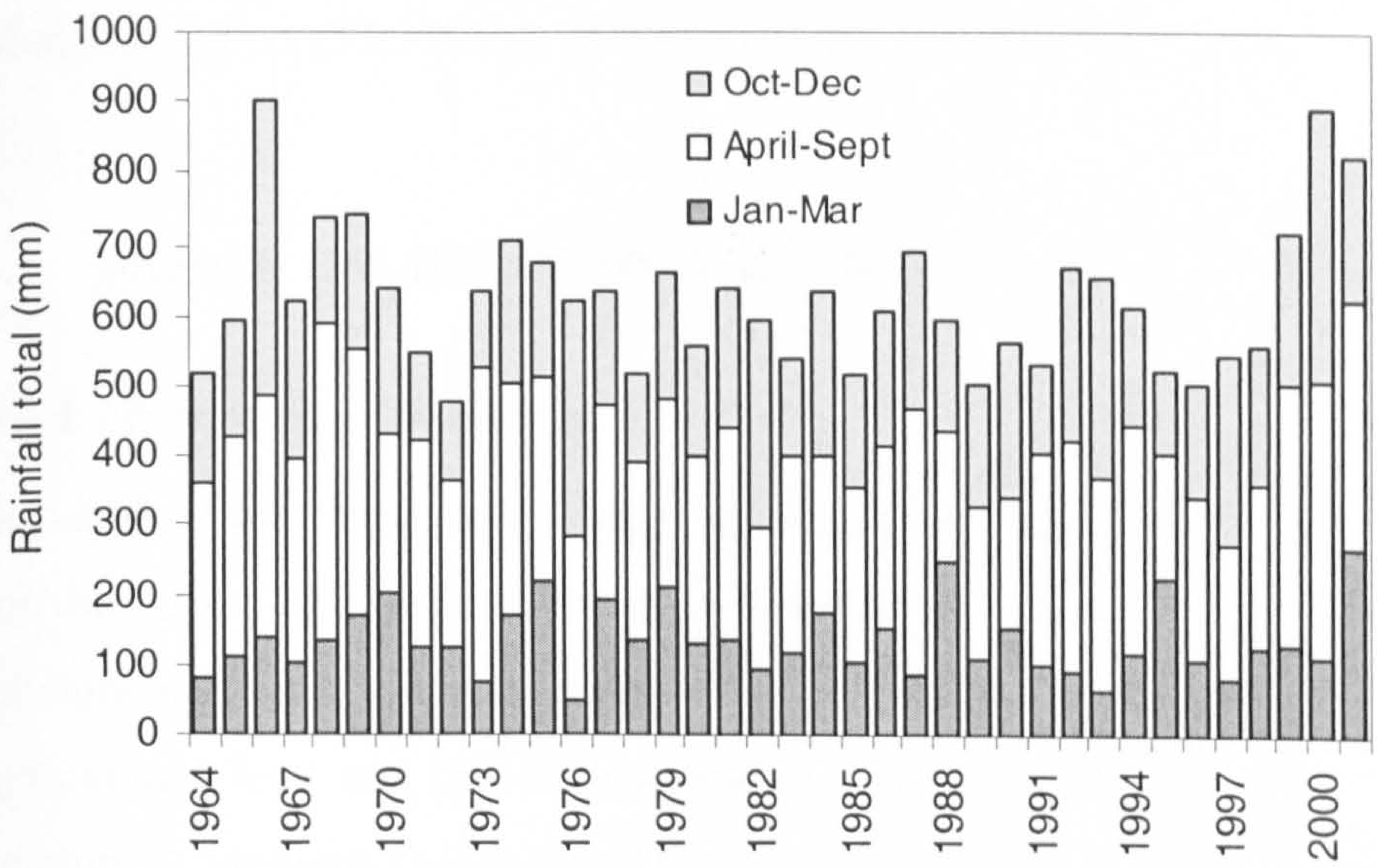


Figure 6.02 – Annual rainfall totals (mm) 1964 – 2001, divided into three periods – April to September representing summer and January to March and October to December representing winter.

Annual rainfall totals range from a minimum of 477 mm in 1972, to the wettest year of 900 mm in 1966. The mean annual rainfall total was 624 mm

6.3.3 Streamflow data

Discharge at the inflow and outflow points of the reserve of the Lampen Stream was measured using a Doppler meter (Starflow, Unidata Ltd.) as described in Chapter 5 (section 5.3.4). The inflow to the site was measured just outside Stodmarsh village within a large rectangular culvert under the road. The outflow measurement site was in a 0.76 m diameter cylindrical pipe under a road. The data was downloaded every month between April 2000 and October 2002. However due to instrument failure, caused by flooding in spring 2001 and further instrument failure in spring 2002 some data were lost.

6.3.4 Evapotranspiration data

Evapotranspiration was measured at Stodmarsh NNR using the Bowen ratio approach during the summers of 2001 and 2002 (see Chapters 3 and 4). The principle objective of this was to calibrate a model to use within the water balance. Modelling was based on the Penman Monteith equation and historical meteorological data from 1973 from Manston airport (51.35°N, 1.35°E) were used as data inputs to this equation.

6.4 MODEL CALIBRATION, VALIDATION AND APPLICATION

6.4.1 Streamflow modelling

Streamflow was modelled using a rainfall-runoff model. A separate model was created for the inflow and outflow points of the reserve, as each of these relates to a different catchment. The inflow catchment is the simpler of the two, consisting almost entirely of agricultural land and therefore a model is more easily created. However the outflow catchment includes the wetland itself, adding complications related to storage on the reserve, making modelling more difficult.

6.4.1.1 *Rainfall input*

Rainfall data is from surrounding raingauges. Although the Stodmarsh National Nature reserve is within the West Stourmouth and Westbere Thiessen polygons, when modelling stream flow at the inflow and outflow of the site, Thiessen polygons must be drawn for the appropriate Lampen Stream catchment. This results in additional raingauges becoming important. Due to the non-continuous availability of five out of the seven raingauges, the Thiessen polygons were redrawn for each of the six combinations of available gauges. The Thiessen weightings for the inflow and outflow catchments are in Tables 6.03 (inflow) and 6.04 (outflow).

Table 6.03 – Thiessen polygon weightings for the inflow catchment. The grey areas indicate that no data were available.

	West Stourmouth	Garrington	Canterbury	Elmstone	Westbere	Bekesbourne	Sturry
1964-67	0.12		0.88	0.00			
1967-69	0.00		0.38	0.00	062		
1969-86	0.00	0.31	0.22	0.00	0.48		
1986-94	0.07	0.67	0.26	0.00			
1995-98	0.07	0.67	0.26				
1998-02	0.05	0.62	0.24			0.07	0.02

Table 6.04 – Thiessen polygon weightings for the outflow catchment. The grey areas indicate that no data were available.

	West Stourmouth	Garrington	Canterbury	Elmstone	Westbere	Bekesbourne	Sturry
1964-67	0.45		0.55	0.00			
1967-69	0.29		0.24	0.00	0.47		
1969-86	0.29	0.19	0.13	0.00	0.38		
1986-94	0.42	0.41	0.16	0.00			
1995-98	0.42	0.41	0.16				
1998-02	0.41	0.39	0.15			0.05	0.01

Where missing data existed within each section, the same approach as described above was used, employing regression equations with the station with which the station with the missing data has the strongest correlation as measured by the r^2 value.

6.4.1.2 Calibration and validation data

The observed data set was split into two to allow for separate calibration and validation periods. However due to the overall limitation in data collection time imposed by the nature of the research the calibration period was much longer than the validation period in order to create a better model. Inflow was calibrated against data from 18/04/00 – 29/09/01 (529 days) and validated against data from 16/07/02 – 07/02/02 (84 days). Outflow was calibrated against data from 30/06/00 – 16/07/02 (746 days) and validated against data from 13/05/00-29/06/00 (47 days).

6.4.1.3 Goodness of fit statistics

Goodness of fit statistics were calculated to assess how well the modelled data fitted the observed data. A number of tests commonly used in hydrology were applied following Aitken (1973), as it is important to establish the presence of both random and systematic errors.

Coefficient of Determination (D_c):

$$D_c = \frac{\sum(q_c - \bar{q}_c)^2 - \sum(q_c - q_{est})^2}{\sum(q_c - \bar{q}_c)^2} \tag{6.02}$$

where q_c is observed runoff, $\bar{q}_c$ is mean observed runoff and q_{est} is estimated runoff from regression line of q_c on q_e and q_e is estimated runoff. D_c is always less than 1 and the closer to 1, the better the model. This test measures the degree of association but does not reveal systematic errors.

Coefficient of Efficiency (E):

$$E = \frac{\sum(q_c - \bar{q}_c)^2 - \sum(q_c - q_e)^2}{\sum(q_c - \bar{q}_c)^2} \tag{6.03}$$

E is also less than 1. If the results are highly correlated but biased, the value of E will be lower than that of D_c.

Sign tests are a simple method of testing whether there are systematic errors. A plus sign was allocated to each overestimate and a negative sign to each underestimate. A chi-squared test was used to compare the number of plus and minus signs with the number expected in an even distribution.

The residual mass curve was computed by subtracting the mean monthly flow from each individual flow to obtain the residuals, which were summed. The ordinate of the residual mass curve depends on the history of preceding events. Comparisons of residual mass curves for observed and estimated flows may therefore reveal the existence of systematic errors in the estimated flows. The residual mass curve coefficient (R) measures the association between the observed and estimated mass curves:

$$R = \frac{\sum(D_o - \bar{D}_o)^2 - \sum(D_o - \bar{D}_e)^2}{\sum(D_o - \bar{D}_o)^2} \tag{6.04}$$

where D_o is departure from mean for observed residual mass curve and D_e is departure from mean for estimated residual mass curve. This statistic is better than D_c or E because it measures the relationship between the sequence of flows and not just the relationship between individual flow events. This should indicate systematic errors.

6.4.1.4 IHACRES model

6.4.1.4.1 Introduction

The streamflow into and out of the reserve was modelled using the IHACRES (Identification of unit Hydrographs and Component flows from Rainfall Evaporation and Streamflow data) model. The IHACRES model is based on the unit hydrograph approach, which uses the concept of the response of the catchment - a unit of runoff production – over a particular time step. It is a lumped, PPCM model at the catchment scale. It was developed by a collaboration between the Centre for Ecology and Hydrology, Wallingford and the Centre for Resource and Environmental Studies in Canberra, Australia (Jakeman *et al.* 1990). The model attempts to avoid the problem of hydrograph separation found in classical unit hydrograph models by relating total rainfall to total discharge. The data requirements are rainfall data, temperature data and observed streamflow data. The model assumes that there is a linear relationship between effective rainfall and runoff produced and that there is a homogeneity of infiltration capacity, rainfall amount and rainfall intensity throughout the catchment. This homogeneity does not exist in reality but as most variation cancels out due to averaging, the procedure can reproduce streamflow with convincing accuracy (Jakeman *et al.* 1990).

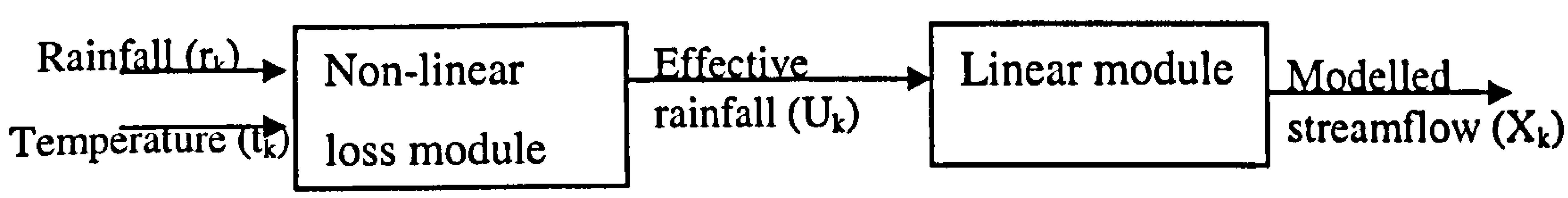
IHACRES has been used to study a wide variety of catchments and has the advantage of a small number of parameters and high predictive accuracy (Hansen *et al.* 1996). It has been shown that the parameters can describe the essential dynamics of the hydrological response of the catchment and are related to physical catchment attributes. An advantage of IHACRES is that the data are allowed to suggest the form of the transfer function used. IHACRES has "the right sort of functionality to reproduce hydrological responses at the catchment scale and about the right number of parameters to be identifiable given a period of calibration data, at least for some environments" (Beven

2000 p.90). It needs no more data than other techniques, baseflow is part of the model – it does not need to be known in advance and time doesn't need to be spent selecting good storm peaks – instead the whole data set is used (Jakeman *et al.* 1990).

Chiew *et al.* (1993) compared six rainfall models of varying complexity including IHACRES in five catchments in Australia. The models were compared using sums of squares and weighted sums of squares to account for low flows. They found that a complex model with 17 parameters performed best in all simulations but IHACRES – significantly less complex – was second best on both a daily and monthly basis. On a daily basis IHACRES was deemed to have unacceptable performance ($E < 0.6$) and this was mainly due to poor daily low flow simulations. However on a monthly basis it was acceptable and the values of the objective function are within 20% of the best model in four out of five wet catchments. Overall IHACRES was deemed an acceptable model in wetter catchments. Hansen *et al.* (1996) used IHACRES to extend historical streamflow records in eight catchments of varying sizes in Australia. On a daily basis five out of the eight catchments were good or very good. An average or good stream gauge rating is capable of producing top performance providing raingauge density is accurate

6.4.1.4.2 Methodology

IHACRES comprises 2 modules in series:



The first module is a non-linear rainfall filter which estimates effective rainfall, the part of rainfall that leaves the catchment as streamflow. It takes account of short term variations, such as soil moisture content and of longer term variation such as evapotranspiration. The second module is the linear relationship between effective rainfall and streamflow. Temperature can be used as an index of evapotranspiration but this may not be necessary if the calibration period is short. The model works on timesteps (k) which represent one day. Effective rainfall is:

$$u_k = r_k \frac{(s_k + s_{k-1})}{2} \tag{6.05}$$

where u_k is effective rainfall (mm), r_k is daily rainfall (mm) and s_k is a catchment wetness index where:

$$s_k = c_{rk} + \left(1 - \frac{1}{\tau_w(T_k)}\right) s_{k-1} \tag{6.06}$$

where C is a volume forcing constant, T is air temperature (°C) and τ_w is a catchment drying time constant where:

$$\tau_w(T_k) = \tau_w \exp(0.062 f (T_{ref} - T_K)) \tag{6.07}$$

where T_{ref} is reference temperature (°C) and f is a temperature modulation factor. $\tau_w(T_k)$ was set as a constant in the inflow model as using varying temperature did not improve the model. However this equation was used with temperature data for the outflow model. C is calculated by IHACRES so that the sum of the observed flow is equal to the sum of the modelled flow.

The second module is the linear relationship between effective rainfall and streamflow. There are three types of model that can be used, representing different combinations of linear storage (Figure 6.03).

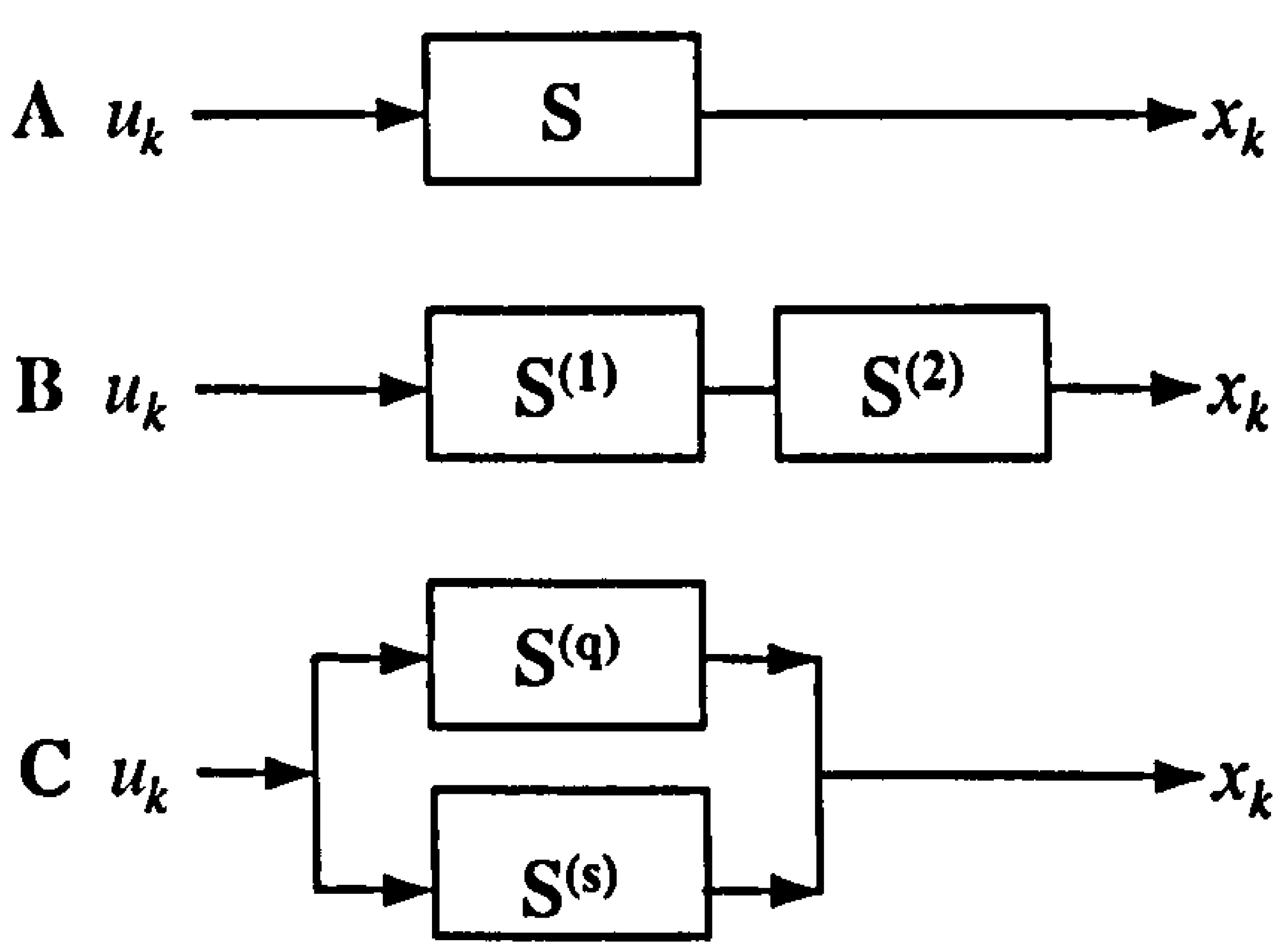


Figure 6.03 – The three configurations of linear storage available in IHACRES. The diagram shows relationships between effective rainfall, u_k, and modelled streamflow, x_k. (Littlewood 2001)

Option A is a first order transfer function, and B and C are second order transfer functions. For the inflow model the second order transfer function was used, with which it is possible to divide the flow into quick and slow flow (option C). Quick flow is mainly the storm hydrograph and slow flow the recession between storms. The equation is:

$$Q_k = \left(\frac{b_0 + b_1 z^{-1}}{1 + az^{-1} + a_2 z^{-2}}\right) u_k \rightarrow Q_k = b_0 U_k + b_1 U_{k-1} - a_1 Q_{k-1} - a_2 Q_{k-2} \qquad (6.08)$$

where Q_k is streamflow output, a_1, a_2, b_o and b_1 are constants and z^{-1} is the backwards shift operator which gives the timestep so:

$$b_1 z^{-1} = b_1 u_{k-1} \qquad (6.09)$$

For the outflow model a first order transfer function (option A) was used with a time delay in the rainfall data of two days. The fact that IHACRES could only identify a single storage component is consistent with the increased routing effect of the ponding of water on the reserve. This suggests that the storage of water on the reserve is dominating the hydrological response of the catchment up to the outflow point, whereas in the inflow catchment dominant quick and baseflow components were identified (Littlewood, personal communication 2002). The outflow equation is:

$$Q_k = \left(\frac{b_o}{1 + a_1 z^{-1}}\right) u_k \rightarrow Q_k = b_o u_k - a_1 Q_{k-1} \qquad (6.10)$$

The *a* and *b* parameters were calculated in IHACRES using the instrumental variable technique. The best model is selected on the basis of a trade off between the coefficient of determination (D_c), which is the proportion of initial variance in streamflow accounted for by the model in calibration mode, and percent average relative parameter error (*% ARPE*), which describes the precision in the *a* and *b* parameters.

6.4.1.4.3 Calibration

The constants were calculated separately for the inflow and outflow points by Ian Littlewood at CEH Wallingford using the IHACRES software, as it is not available for general distribution (Table 6.05).

Table 6.05 – Calibrated constants used in the IHACRES models

Constant	Inflow	Outflow
C (mm^{-1})	0.00349	0.024135
$\tau w(tk)$	34 days	
τw		18
f		1.7
R	20°C	0°C
a_1	-1.684	-0.968056
a_2	0.687	
b_o	0.00799	0.003469
b_1	-0.00775	

These constants were set up within the above equations in a spreadsheet in order to use the model and the calibrations can be seen in Figure 6.04. The inflow calibration includes a period of missing data which had to be interpolated first.

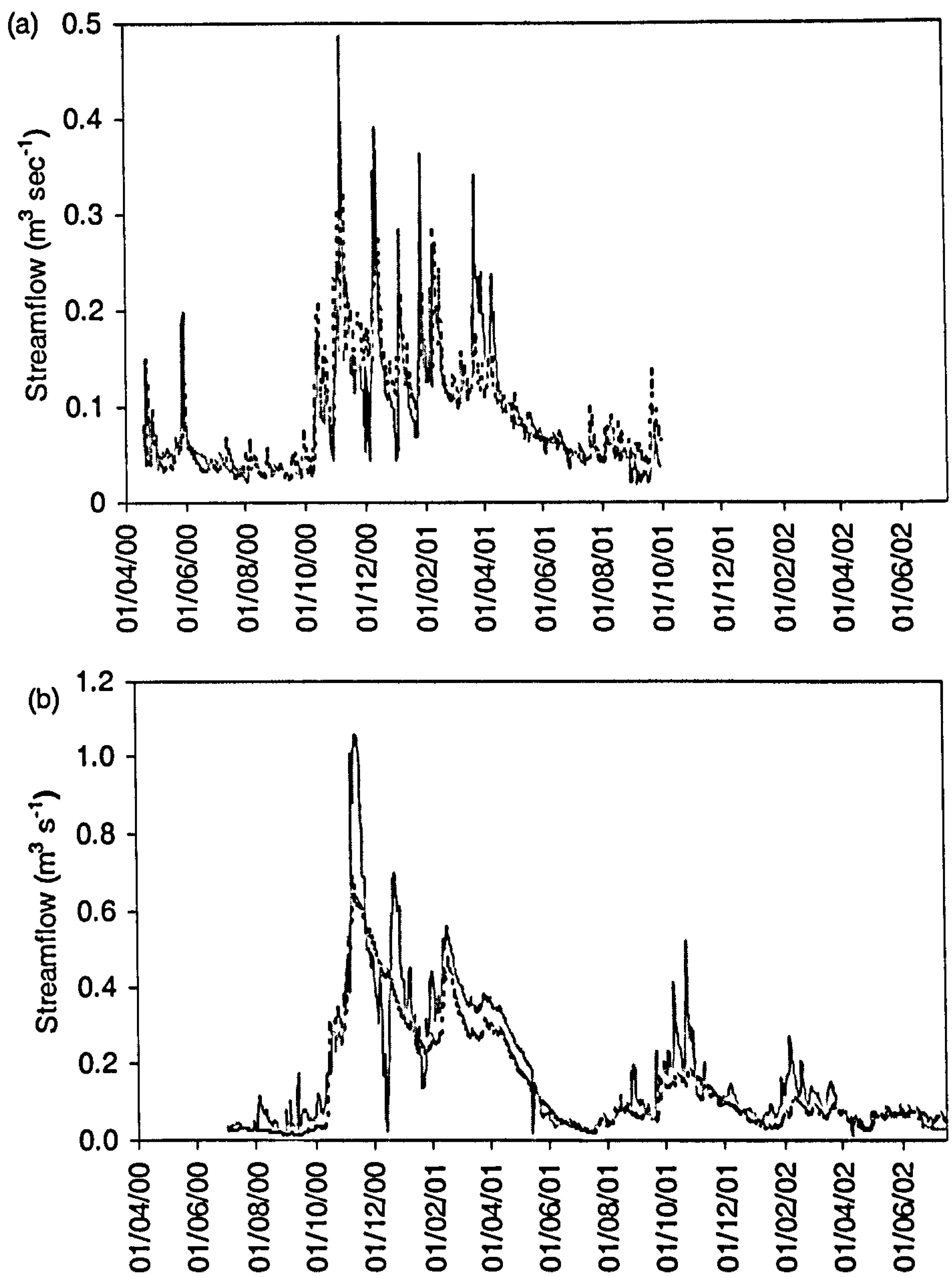


Figure 6.04 – Calibration of IHACRES modelled (a) inflow and (b) outflow against observed streamflow. The grey solid lines show observed data and the black dashed lines modelled data.

6.4.1.4.4 Validation

Both the inflow and outflow models were validated using separate data sets from those that were used in the calibration. However in order to maximise calibration data only small sets of data remained for validation.

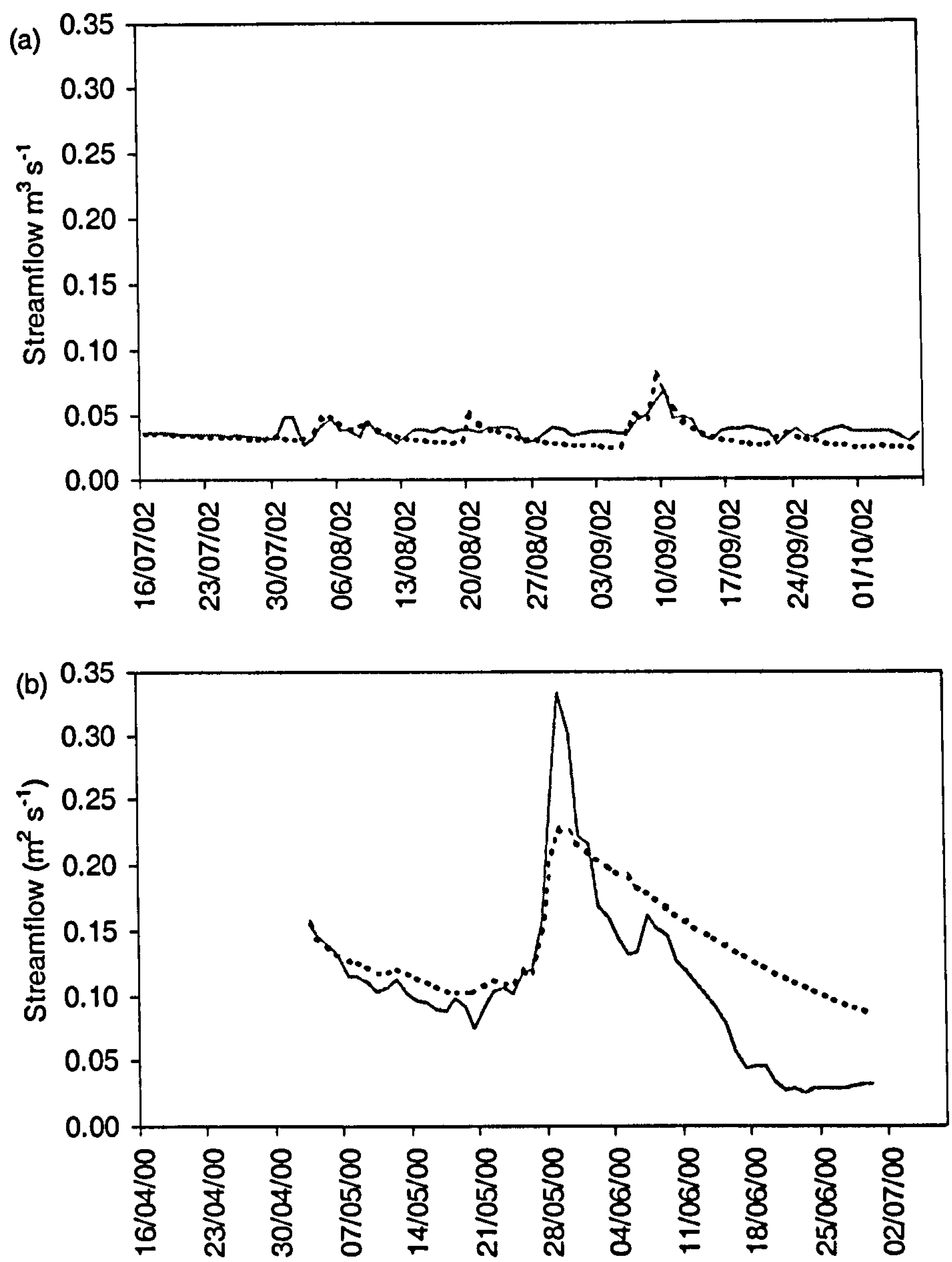


Figure 6.05 – Validation of predicted IHACRES (a) inflow and (b) outflow against observed streamflow. The grey solid lines show observed data and the black dashed lines modelled data.

The validation period used for the inflow was over a year after the end of the calibration period. The validation period is a period of low flow with little change in streamflow discharge though it can be seen that the one small peak is captured quite well. Although the outflow validation starts well, the model output does not reach the peak of the storm response and does not fall away quickly enough.

The results of the goodness of fit tests are shown for the IHACRES model for both calibration and validation in Table 6.06. The goodness of fit tests are rated following Chiew *et al.* (1993) who found that a coefficient of 0.9 or above indicated a perfectly acceptable model and 0.6 or above an acceptable model.

*Table 6.06 – Goodness of fit test statistics for calibration of the IHACRES model. ** - perfectly acceptable model, * . acceptable model. No asterisk – unacceptable model (following Chiew et al. 1993)*

	Inflow		Outflow	
	Calibration	Validation	Calibration	Validation
D_c	0.951 **	0.77 *	0.901 **	0.54
E	0.772 *	0.55	0.790 *	0.45
X^2 of sign test	18.08	7.7	47.00	11.6
R	0.912 **		0.948 **	

For both models the calibration value for D_c is close to one, indicating that there is a high degree of association between the modelled and observed values. E is lower however, which may indicate that there is some systematic error, confirmed by the high values of the X^2 of the sign test. R, however, which is recommended as the best test for systematic error, has a high value. The residual mass curve coefficient could not be calculated for the validation data as this uses monthly data and the validation data sets were too short. The outflow validation statistics are low compared to the calibration indicating potential problems with the model. The inflow model shows a good value for the coefficient D_c. The coefficient E is lower because of the lack of variation in the data from the mean.

6.4.1.5 Alternative outflow models

It was decided that the IHACRES inflow model was acceptable. However due to the poor validation of the outflow model some alternative modelling approaches were tried. The details of the models and figures describing in detail their calibration and validation can be found in Appendix H. Neither of the alternative models resulted in any improvement on the IHACRES model, either in calibration or validation.

6.4.2 Evapotranspiration modelling

Evapotranspiration is a difficult parameter to model and this problem has been considered in Chapters 3, 4 and 5. There are two main approaches that have been considered:

1. The crop coefficient approach, using coefficients either from (a) the Bowen ratio approach, (b) from the work of Fermor *et al.* at their Walton lake site (the site most similar to Stodmarsh) (2001) or (c) the standard 1.4 coefficient as suggested by Bardsley *et al.* (2001) and Hawke and Jose (1996)
2. Using the directly calibrated Penman Monteith equation that was developed for reeds in Chapter 4.

There are potential problems with both approaches. Both result in a lot of scatter in calibration. We have no coefficients from the Bowen ratio approach for the winter months though the water balance showed this to be fairly unimportant.

A number of different sets of crop coefficients were compared with each other and with the direct Penman Monteith method (Table 6.07).

Table 6.07 – Crop coefficients compared in the water balance model

Month	Current value	Fermor (Walton Lake)	Bowen ratio	Bowen ratio with rain days dry / wet
January	1.4	1.09	0.75	0.60 / 0.85
February	1.4	0.46	0.75	0.60 / 0.85
March	1.4	0.48	0.75	0.60 / 0.85
April	1.4	0.78	0.75	0.60 / 0.85
May	1.4	0.63	0.75	0.60 / 0.85
June	1.4	0.77	0.73	0.67 / 0.89
July	1.4	0.86	0.68	0.61 / 0.85
August	1.4	0.72	0.78	0.69 / 0.92
September	1.4	0.75	0.75	0.60 / 0.85
October	1.4	0.82	0.75	0.60 / 0.85
November	1.4	0.76	0.75	0.60 / 0.85
December	1.4	0.97	0.75	0.60 / 0.85

The models were compared to see which fitted best into the water balance models of 2000 and 2001, creating the best fit between measured and predicted change in storage data. The water balance models were created with measured streamflow data. This method of validation was chosen as the aim is to create a model with as accurate a prediction of change in storage data as possible. For this reason the crop coefficients back calculated from the water balance could not be included. The accuracy of the predicted change in storage data is assessed using the mean sum of squares and the explained variance as suggested by Liden and Harlin (2000):

$$R^2 = 1 - \frac{MSE(Q)}{VAR(Q_{obs})} \qquad (6.11)$$

If the result of this equation is negative then this indicates that the simulation is worse than if average values were used.

Table 6.08 shows the goodness of fit of the predicted storage data to the measured storage data using different evapotranspiration models as assessed using Equation 6.11. Each model is fitted to the summer 2000 and summer 2001 data and the combined data. The winter was not included due to the large unknown component in the water balance.

Table 6.08 – Comparison of evapotranspiration models using an R^2 value as defined by Equation 6.11

	2000	2001	overall
Direct Penman Monteith	-0.79	-3.8	-0.69
Bowen ratio *Kc*	0.684	-0.976	0.195
Bowen ratio *Kc* with rain days	0.698	-9.488	0.23
Fermor *Kc*	0.776	-1.969	0.669
Kc of 1.4	-6.896	-102.434	-10.583

In 2001 all the R^2 values are negative indicating that using the average value of storage would be an improvement on the model. This occurs because there was very little change in the measured storage values over the months being considered in the water balance, therefore the variance is very low. The model which created the most accurate simulation over the two years is used in the overall water balance model. Overall the Fermor *Kc*s appeared to give the best results, followed closely by the use of Bowen ratio *Kc*s. In addition to giving the best results in the objective analysis above, the Walton Lake *Kcs* of Fermor *et al.* were considered suitable for use in the model because they came from a site with similar characteristics to Stodmarsh in terms of the relatively large area of the site, ensuring no influence of advection and the fact that the reedbed was well established (not the case at the other two sites studies by Fermor *et al*). In addition, there are year round coefficients available, whereas with the Bowen ratio approach, Kcs must be estimated for nine out of twelve months. The Fermor *Kcs* were used with Manston Airport meteorological data from 1973-2001.

6.4.2.1 Sensitivity analysis

The measured water balance of 2000 was analysed to determine how sensitive it is to the crop coefficients used within it. The graph below shows the impact of changing the crop coefficients from the Fermor coefficients between 5% and 30%.

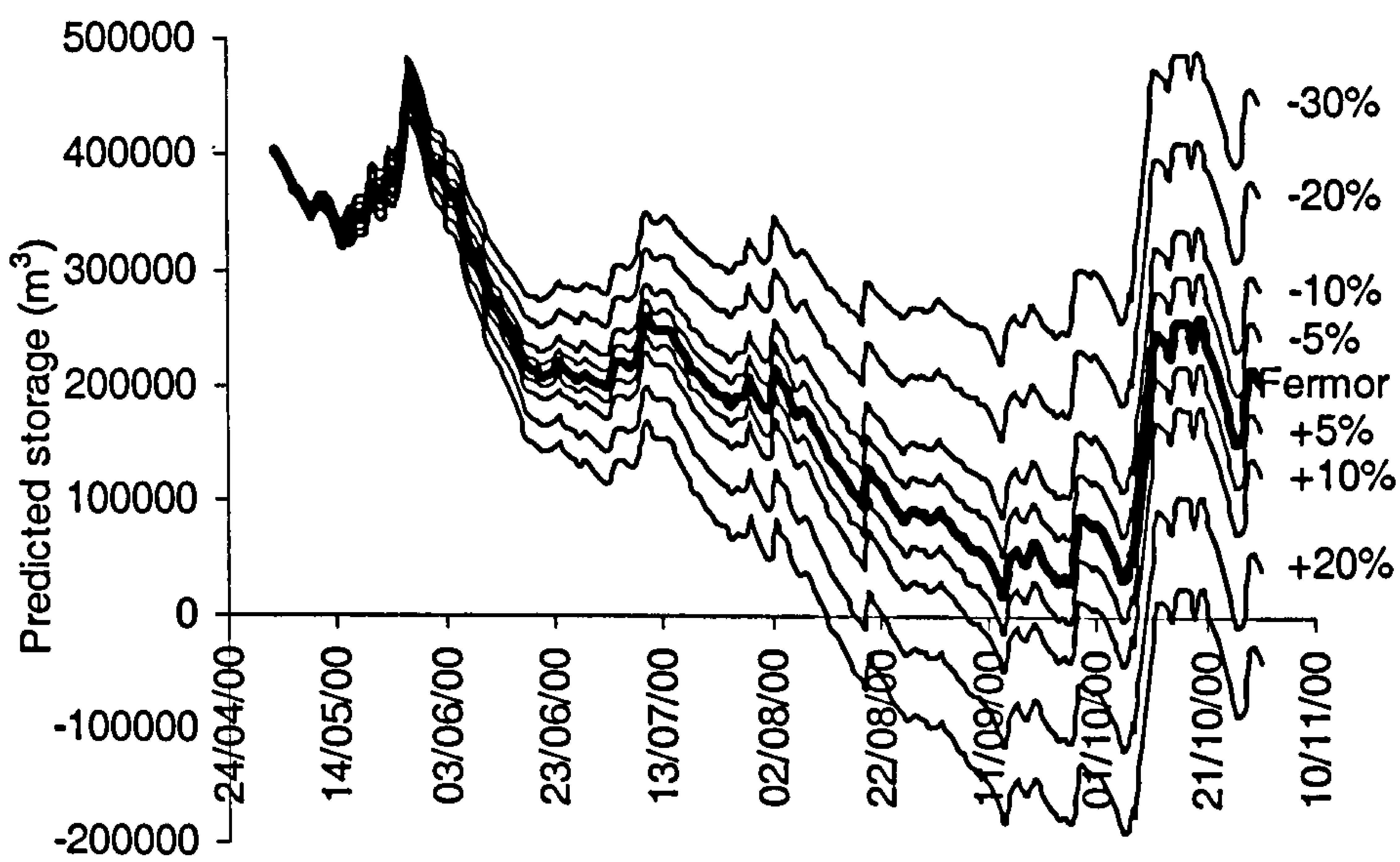


Figure 6.06 – Sensitivity analysis of crop coefficients

It can be seen that close to the beginning of the model the sensitivity to crop coefficients is low. However because the water balance is a cumulative model the error in predicted storage caused by a particular error in crop coefficients increases as time goes on. Table

6.09 shows the percentage change in predicted storage, compared with that predicted using the Fermor coefficients on the first of each month (the model began on 3rd May)

Table 6.09 – Sensitivity to crop coefficients

	Percentage change in crop coefficients from Kc_{Fermor}			
	5%	10%	20%	30%
	Percentage change in storage estimation			
1st June	1.28	2.58	5.15	7.73
1st July	6.65	13.3	26.6	39.9
1st August	11.92	23.84	47.69	71.53
1st September	35.46	70.92	141.83	212.76
1st October	43.98	87.95	175.90	263.86

Errors in the model formulation and parameterisation are compounded as run time is increased.

6.4.3 Overall model

The overall model was created using Equation 6.01. The model used 29 years of daily historical rainfall data and calculated historical evapotranspiration data. The IHACRES inflow and outflow models were used to estimate streamflow. Each day the change in storage was calculated from the water balance and the cumulative change found on a seasonal and annual basis. On an annual basis, the overall change in storage between the beginning (when water levels were set) and end of the year was found as the surplus or requirement for water on the site for that year.

The water requirements of the site are that the water should be at 200 – 300 mm in March. However if a single abstraction is to occur it will take place around the end of December. According to the model, the mean change in storage between January and March is an increase of 87 mm. Therefore the desired value for the beginning of January was set at 250 mm – 87 mm = 163 mm (395927 m^3 of water over the whole site). The difference between this figure and the storage level on 31st December of the same year was used to calculate the surplus or deficit of water that year. If there was a deficit, this is the amount that should be abstracted in order to regain target water levels for the next year. This surplus or deficit was then related to annual rainfall as shown in Figure 6.07 below.

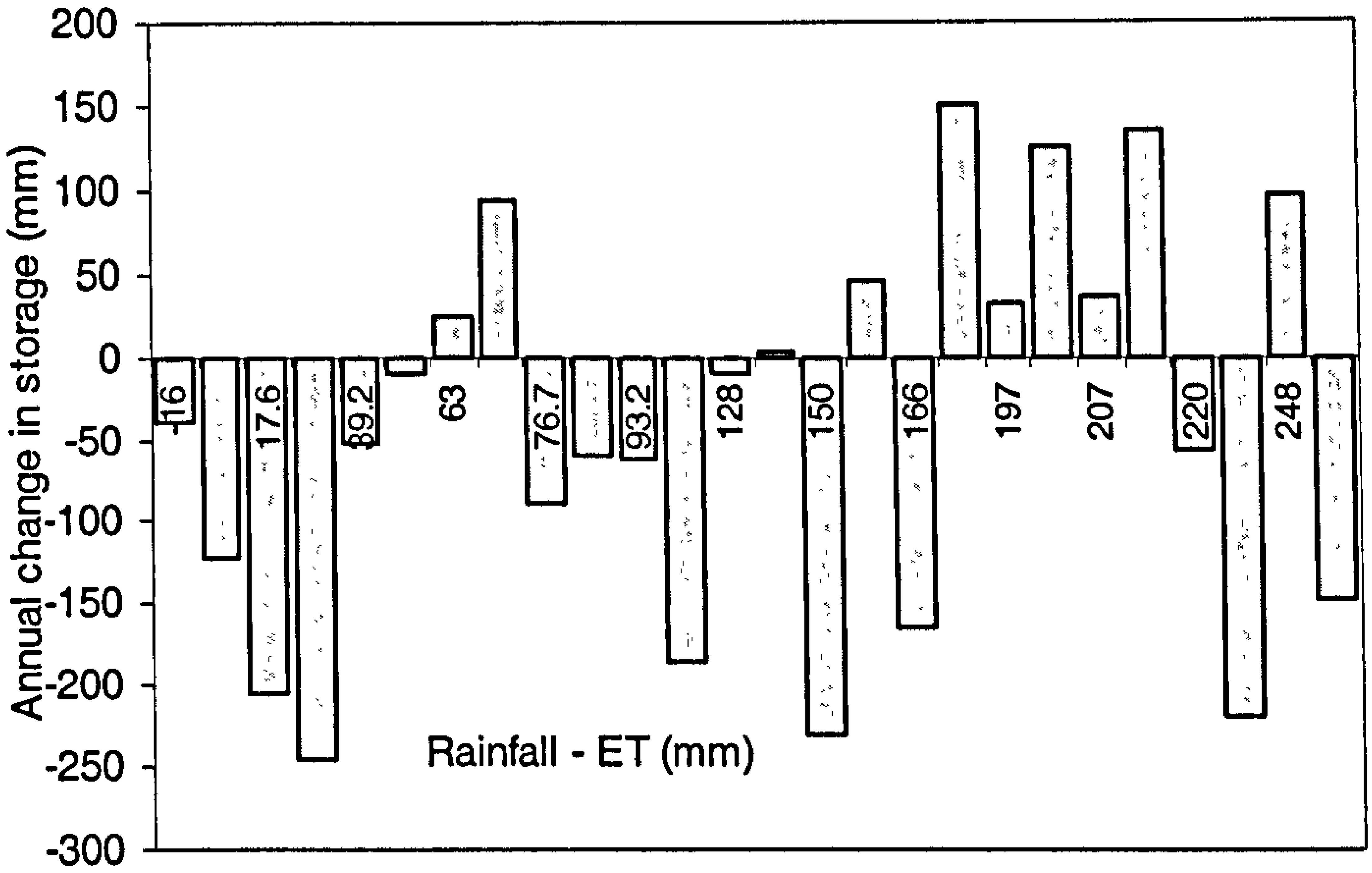


Figure 6.07 – Annual surplus and deficit of water arranged in ascending order of annual total rainfall minus ET.

It can be seen that there is no relationship between total on site precipitation and change in storage. This is intuitively wrong, as experience has shown that there is a clear relationship between dry years and water deficits on site. Further investigations were made on a seasonal basis. Figure 6.08 shows the relationship between rainfall and the change in storage for each year for the period January to March.

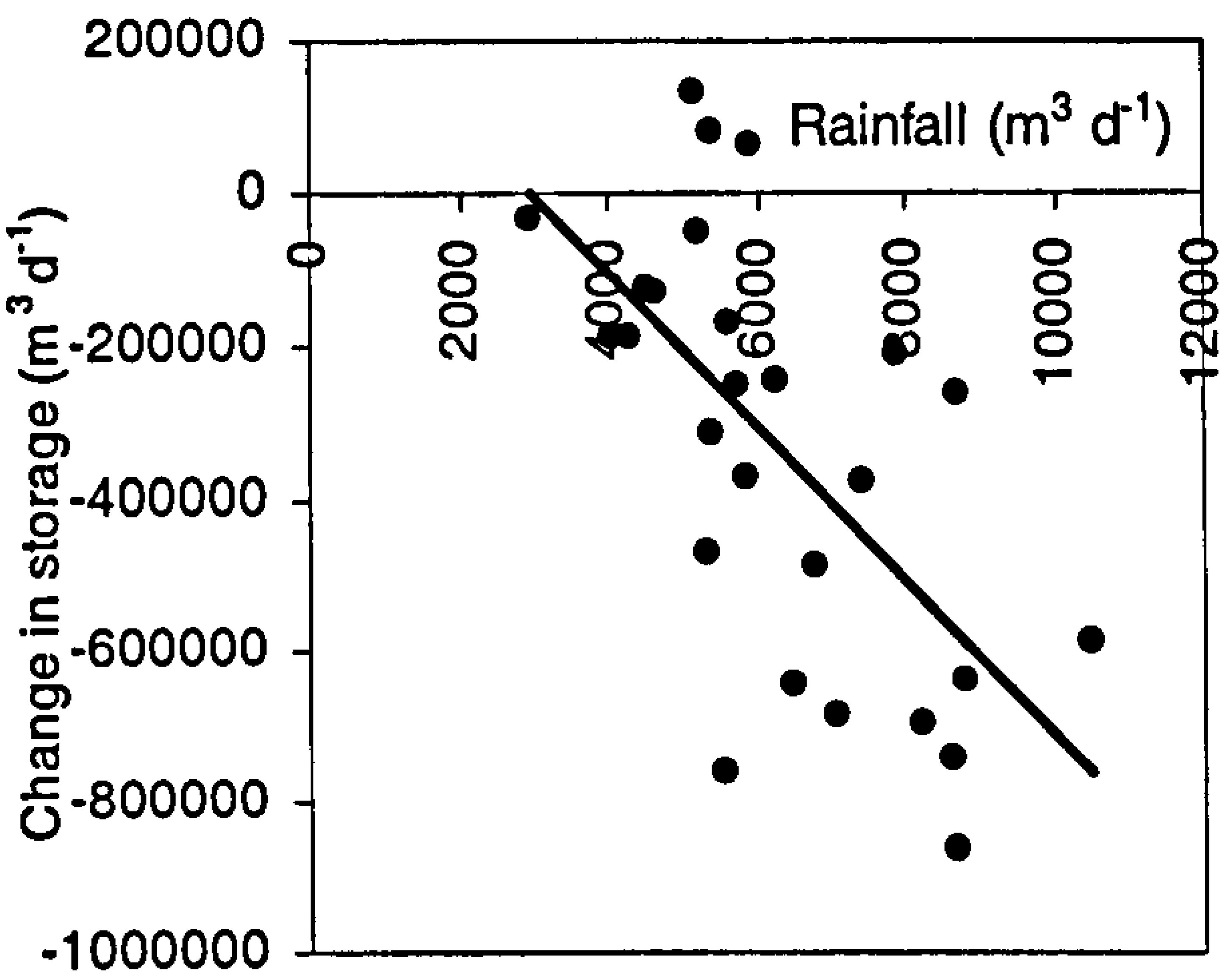


Figure 6.08 – The relationship between rainfall and change in storage between January and March.

The graph shows that the higher the rainfall, the greater the reduction in storage, i.e. wetter weather appears to result in a dryer site. It was found that this was due to over-prediction by the IHACRES outflow model. The IHACRES outflow model responds to rainfall with a large consequential increase in flow out of the site. This is however not matched by a corresponding increase in inflow or even rainfall and consequently outflow exceeds inflow and water appears to be draining from the site. The reason for this problem is that the model was calibrated over a period that was much wetter than average. Much of this time included additional input to the site from the river, which is included in measured outflow. The calibrated model therefore accounts for these extra sources of water beyond the immediate rainfall input resulting in high outputs, not balanced by inputs. In drier times these extra water sources will not exist and therefore the model becomes inaccurate in accounting for them. It can be seen in the validation of the model (Figure 6.05) that it over-predicts measured data. It was concluded that for this reason the IHACRES outflow model could not be used. Alternative outflow modelling approaches were tried and the results are presented in Appendix I. However these models provided no improvement in the level of confidence with which the outflow was modelled. Instead an alternative modelling approach (described below) was attempted which did not require use of the outflow model.

6.4.4 Water requirement modelling approach

All wetlands depend ultimately on precipitation to offset evapotranspiration and outflow losses. There are many wetlands in the UK, particularly in the western uplands, such as blanket peat and raised bogs, which are sustained solely by precipitation. Elsewhere wetlands need additional supplies of water to maintain saturation throughout the summer. The alternative modelling approach described below was based on investigating the rainfall-evapotranspiration deficit at Stodmarsh NNR, and then comparing this deficit with available water supplies from the Lampen stream. This simple water balance was calculated on an daily basis, simply looking at the site as if rainfall were the only input and evapotranspiration the only output. The resultant cumulative change in storage was found on an annual and seasonal basis. Water levels were set at 206 mm on 1st April to minimise cumulative errors. Water levels were allowed to increase on the reserve until a maximum threshold of 250 mm was reached in summer and 300 mm in the winter (based on the maximum desirable depths for the conservation of Bittern (Newbold and Mountford 1997)) at which point it was assumed water would leave as outflow via the Lampen Stream.

Using the cumulative annual results of the rainfall-evapotranspiration water balance and comparing the levels of storage in April each year, out of 29 years (1973-2001) in only 3 years (1990, 1995 and 1996) do total annual evapotranspiration requirements exceed total annual rainfall, so on an annual basis in nearly 90% of years the site is sustainable without any input from the Lampen Stream. However this hides an imbalance between summer (April to September) and winter (October to March) (Figure 6.09).

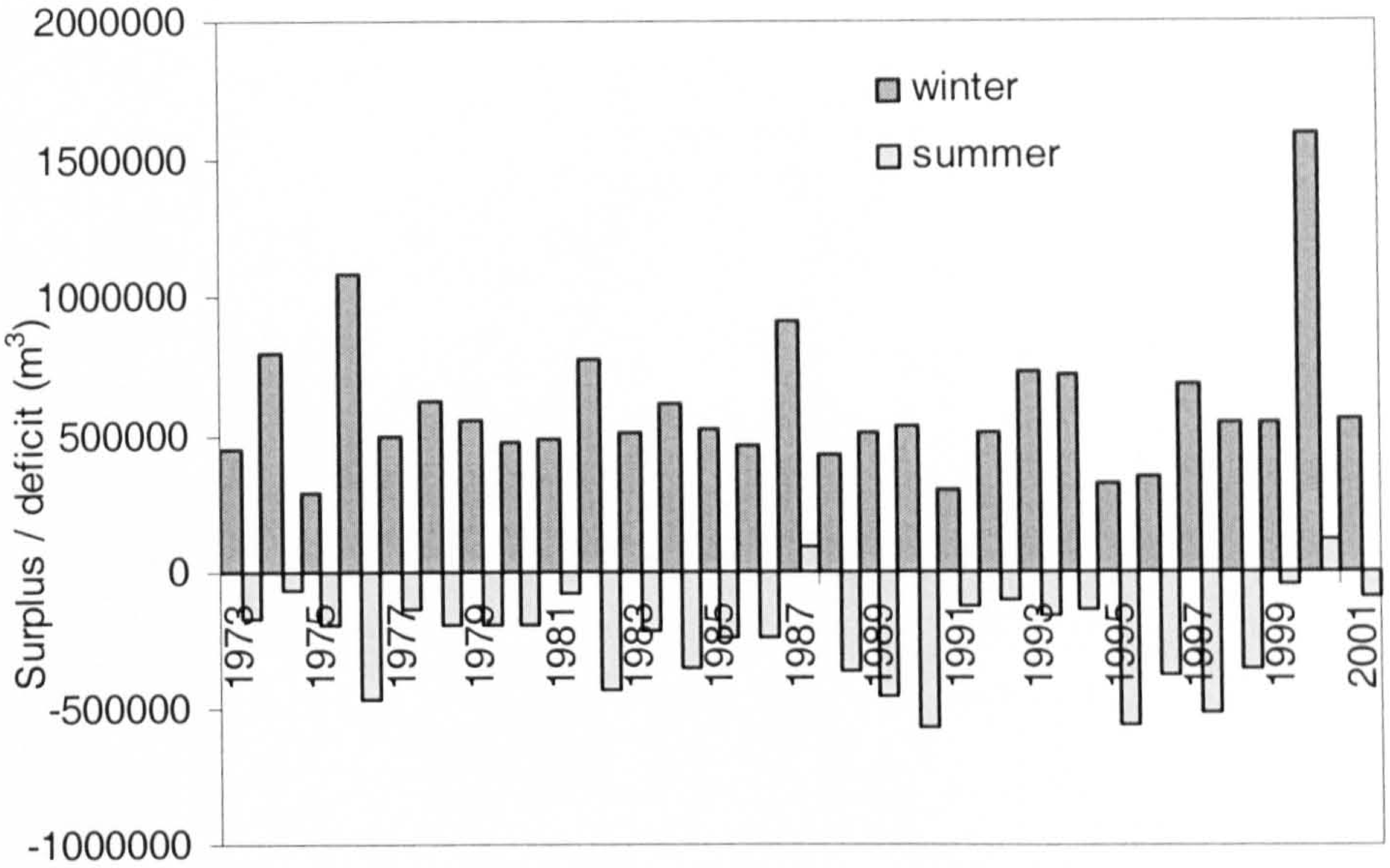


Figure 6.09 – Winter and summer water surplus and deficit based on total rainfall and evapotranspiration volumes.

In almost every year (except 1987 and 2000) there is a deficit of water over the summer months and a surplus in the winter. This is reflected in the water levels on the site. Some examples of annual predicted storage volumes using the water requirements model are shown in Figure 6.10.

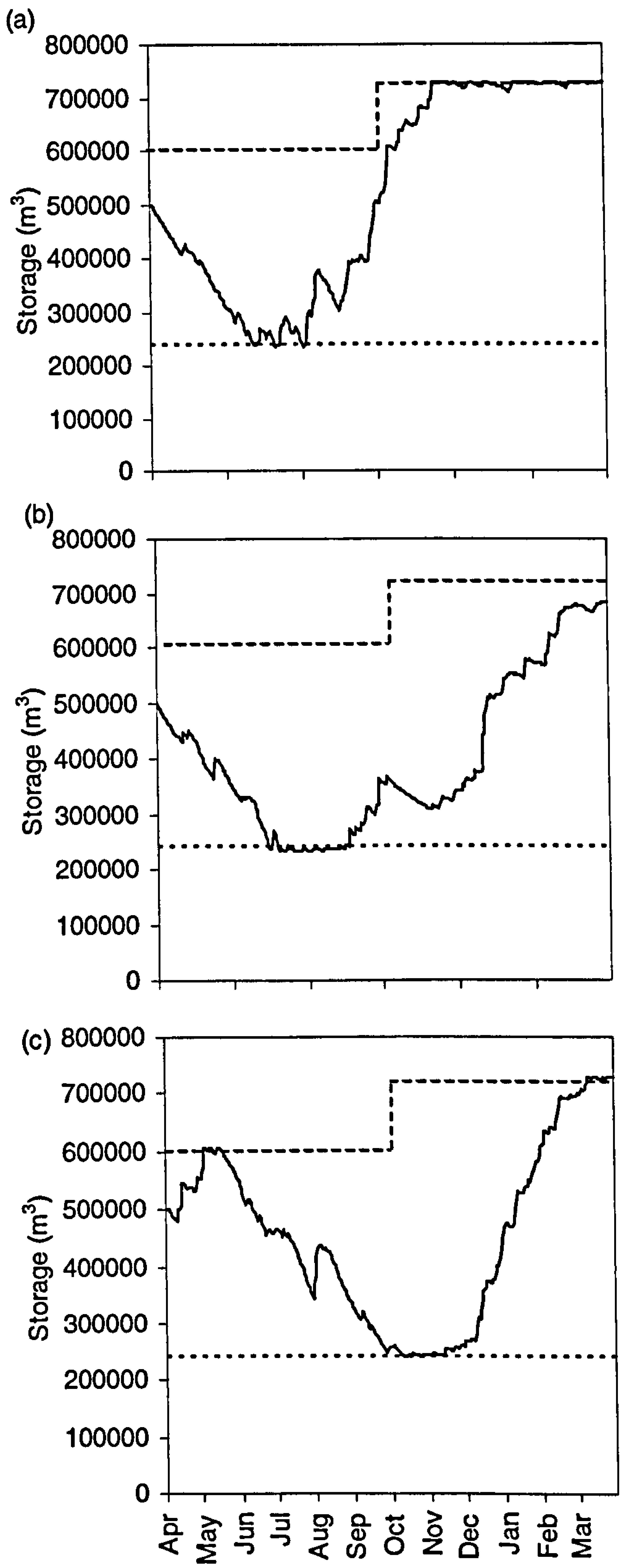


Figure 6.10 – Charts showing the predicted storage on the site based on rainfall and ET in (a) 1974-75 – a wet year, (b) 1994-95 – a dry year and (c) 1977-78 – an average year. The lower dashed line represents a mean depth of 100 mm, the minimum desirable

water level on the site, and the upper dashed line the upper threshold of water on the site.

The prevention of excessive moisture deficits in summer, which would encourage succession from wetland to scrub, is the principle aim of water level management on wetland reserves. It is not only important to have sufficient water to maintain the site on an annual basis, but it is also important to maintain an adequate depth of water on the site throughout the year. The minimum recommended water depth for Bitterns is 100 mm (Newbold and Mountford 1997) and the aim of the Stodmarsh NNR management plan is to have water depths between 100 and 250 mm (Burnham 1999). By impounding and importing water in the months when evapotranspiration demand is highest, the deficit and decline in water levels can be reduced. It was initially assumed that managers can use as much water from the Lampen Stream as they require, using sluices and pen stop dams to prevent Lampen Stream water leaving the reserve. A statement was included in the model that said that should the water level on the site drop below a mean level of 100 mm, water from the Lampen Stream would be used to make up the deficit. Conceptually this water would come initially from natural movement of stream water onto the site using hydrological gradients, and beyond this by abstraction, either by holding back stream water with dams or by using a wind pump. The total amount of water required to make up this deficit each summer was then calculated (Figure 6.11). In 20 out of 29 years some water was required to maintain a minimum mean level of 100 mm.

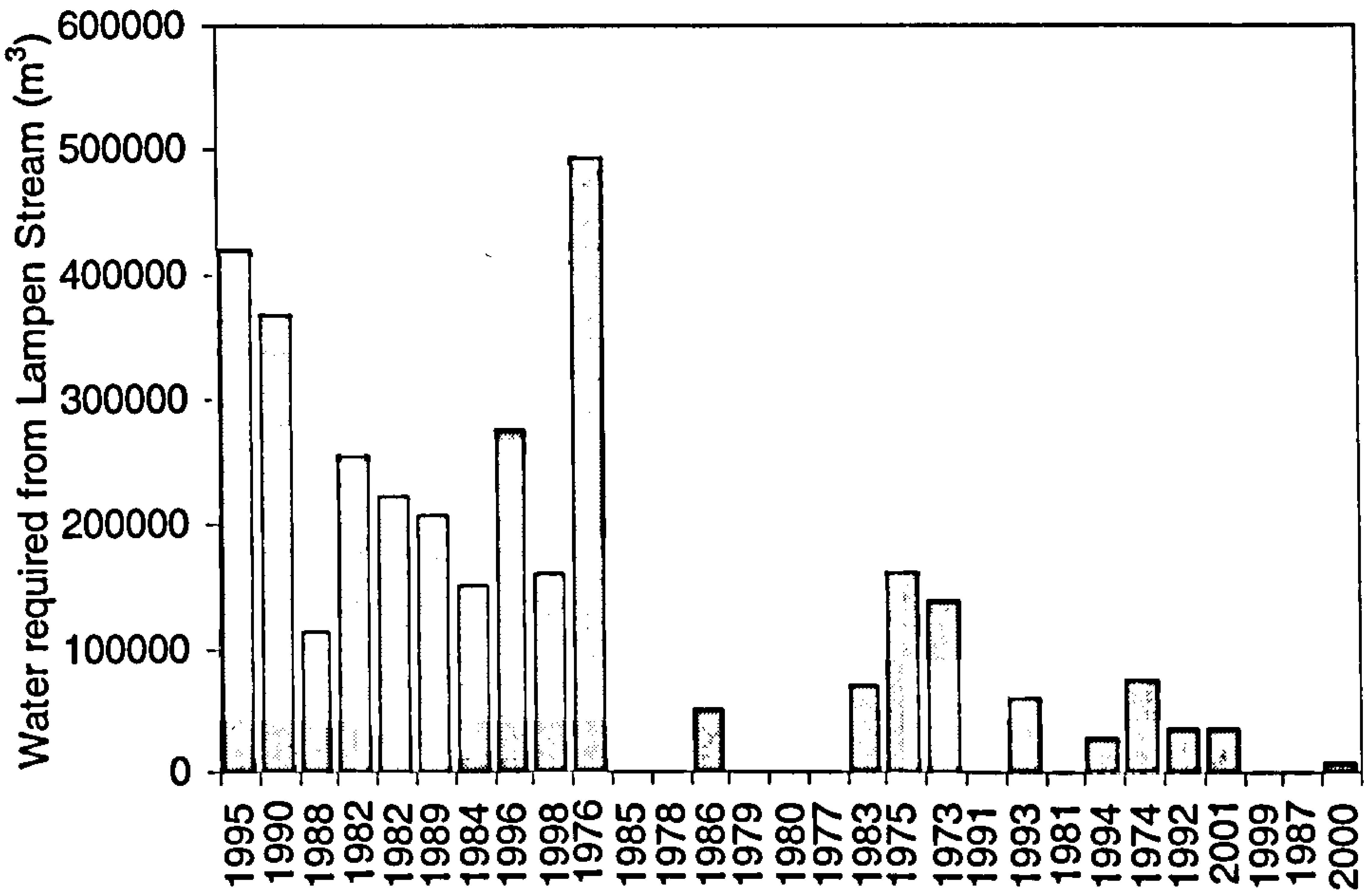


Figure 6.11 – The volume of water required from the Lampen Stream to make up the deficit of rainfall – evapotranspiration in summer arranged in order of increasing summer rainfall.

This can also be expressed as a probability (Figure 6.12). Probability was determined by ranking the years in order of volume of water required and using Equation 6.12:

$$p = \frac{r}{n+1} \tag{6.12}$$

where p is probability, r is rank and n is the number of years in the sample.

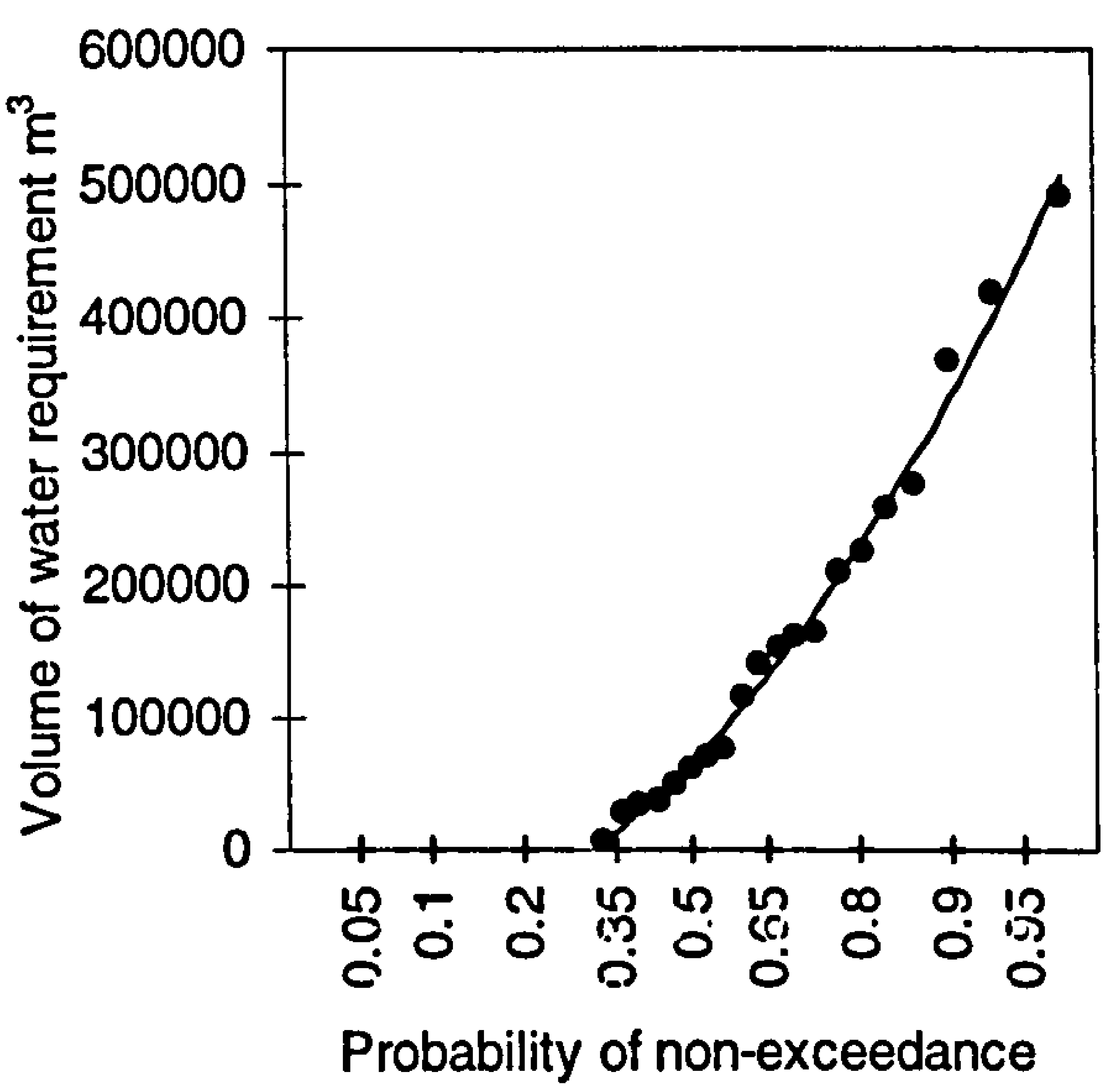


Figure 6.12 – The probability of requiring different volumes of water from the Lampen Stream in order to make up summer deficits.

The next step is to consider water resource availability as predicted by the IHACRES inflow model. Figure 6.13 shows a similar probability distribution of the total predicted discharge in the Lampen Stream over each summer in the 29 year series.

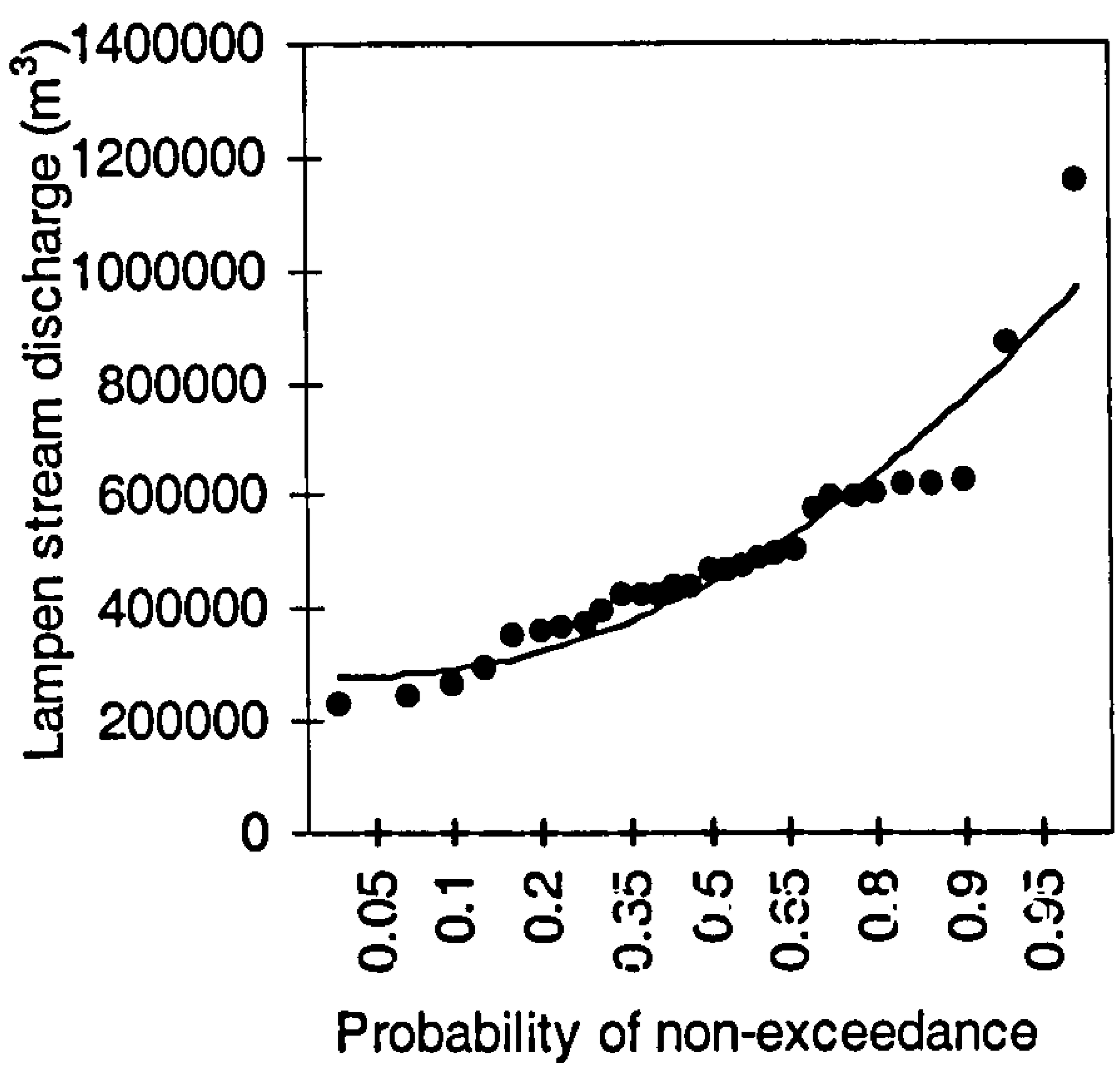


Figure 6.13 – The probability of the reccurence of summer discharges in the Lampen Stream as predicted by the IHACRES model.

In order to investigate whether resource availability is likely to be sufficient to meet requirements, Figures 6.12 and 6.13 can be combined. They cannot simply be overlaid as there is likely to be an inverse correlation between water requirements and water availability – those years where most water is required on the reserve are likely to have the least water available in the Lampen stream. The water required from the Lampen Stream can be calculated as a percentage of water available at the inflow to the site as calculated using the IHACRES model (Figure 6.14).

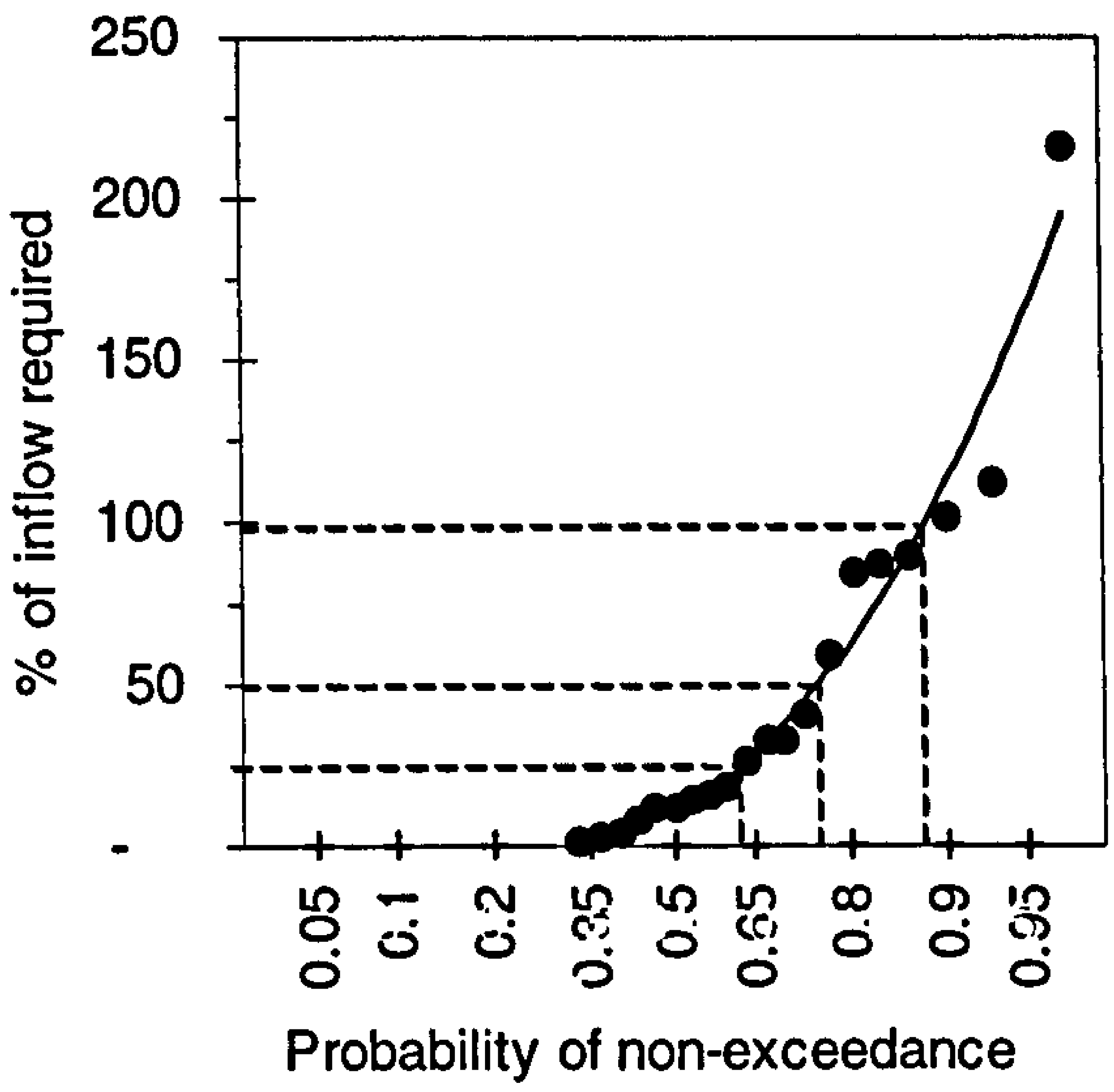


Figure 6.14 – The probability of requiring different percentages of the inflow from the Lampen Stream in order to make up summer deficits. The dashed lines represent the probability of demand exceeding 25%, 50% and 100% of the Lampen stream.

There is a probability of 0.36 that more than 25% of the Lampen Stream discharge is required. In around 27% of years more than 50% of Lampen Stream water will be required over the summer and in around 14% of cases the amount of water required is greater than streamflow available.

Use of Lampen Stream water is limited by the fact that there are other holders of abstraction licences who have a right to water from the Lampen Stream. Water cannot be held back to such an extent that those users have insufficient water to meet their needs. There are three licence holders who can abstract downstream of the point where inflow to the reserve was measured. Between them they can extract 63 138 m^3 water between March and September – the period when it is also most needed by the reserve. As a percentage of the available inflow, this ranges from 5-27%. As the years with the highest demand from abstraction are also likely to be the years with the highest demands from the reserve this could lead to potential conflicts of interest in years of low water availability. The demand for water from the Lampen Stream for the reserve and the maximum potential demand from abstraction were summed and calculated as a percentage of the availability of the Lampen Stream. This is plotted in the probability graph in Figure 6.15.

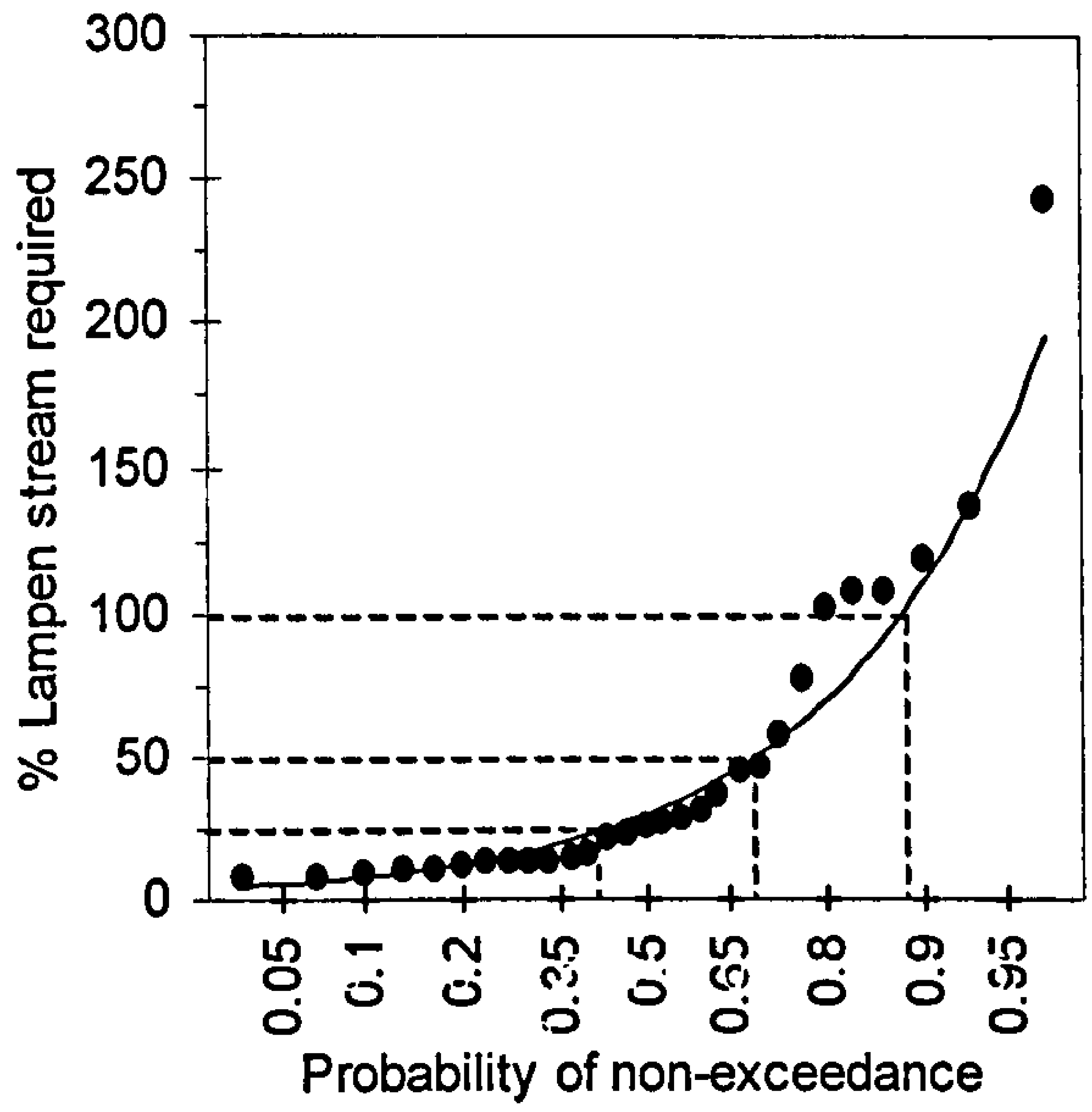


Figure 6.15 – Probability of requiring various percentages of the Lampen Stream to meet the summed demands of Stodmarsh National Nature Reserve and the abstraction of other water users. The dashed lines show the probabilities of needing less than 25%, 50% and 100% of stream water.

Figure 6.15 shows that when abstraction by other water users is included, the number of years where more than 100% of the Lampen Stream water would be insufficient increases to 5 out of 29. In 36% of years more than 50% of the water will be required and in 58% more than 25% will be required.

6.5 DISCUSSION

Bradley (2002) concluded that annual models of wetland hydrology are difficult to develop since there are many potential errors in model formulation and parameterisation and these are compounded as model run time is increased. In addition to the problems of estimating evapotranspiration which have been discussed in detail in previous chapters of this study, the modelling stage has introduced the additional problem of modelling streamflow. Rainfall-runoff models are a frequently used technique and conceptual models such as IHACRES are capable of producing accurate results on well defined catchments. However, ideally, model calibrations require a minimum of three years data, in addition to data for validation. Many studies (e.g. Hansen *et al.* 1996) use calibration and validation periods of five to ten years. Within the time-scale of this research and considering there was no pre-existing flow data for the Lampen Stream meaning that monitoring had to be set up at the start of the work this ideal was an impossible goal. However when this inherent difficulty was compounded by abnormally wet conditions, which led to additional data loss, the data series with which to calibrate the model was far shorter than would be hoped. This also led to compromises in the validation period. For these reasons confidence in the models was lower than it should be. However, reasonable results were found with the inflow model. The coefficient of determination was acceptable by the standards of Chiew *et al.* (1993) and although the coefficient of efficiency was less than acceptable, this was influenced by the small variance in the validation data. Particular problems were experienced with the outflow model however, which includes the water that flows through the reserve. Conceptually this model is more difficult to calibrate, as storage within the reserve becomes important. The stream interacts with the reserve so its discharge is dependent not only simply on soil saturation and rates of through-flow but also on the levels of storage and residence times within the reserve. The fact that the model was calibrated over two out of the three wettest years in the last 37 years meant that the IHACRES outflow model was not calibrated in a way that could be applied to other drier years, despite the fact that its goodness of fit statistics were encouraging in calibration. Boeye and Verheyen (1992) also found this problem when modelling the hydrology of a Belgian groundwater discharge fen using a linear regression model. When the model was validated it was found to be inaccurate because it was calibrated over a wet period and validated over dry. These problems meant that in the current study an alternative modelling approach to the simple water balance equation developed in Chapter 5 was used.

Despite all the work attempting to calibrate an evapotranspiration model for reeds it is disappointing to note that the results in the end were not successful and no improvement could be made on the work of Fermor *et al.* (2001) at Walton Lake. In 2000 both the Fermor coefficients and those developed from the Bowen ratio approach fitted the model well. However in 2001 due to the very small variation in storage over the period used to fit the model, all the goodness of fit coefficients revealed that an average value would give a better result than the changing monthly values. Despite this, Chapter 5 shows that both Bowen ratio results and Fermor did give a reasonable fit. The often-quoted value of 1.4 gave by far the poorest results and using Penman Monteith directly also did not produce successful results. Overall the *Kc*s provided by Fermor *et al.* (2001) were used as they gave the best-fit values over both years and also have year round values available.

As was acknowledged in Chapter 5, the issue of input from the River Great Stour through seepage and flooding over the riverbanks is a major uncertainty in this research. This was known at the start of the work, but as the river input appears to be diffuse it was almost impossible to measure it hydrologically. Chlorine tracing was considered but was impossible on the grounds that chloride concentration was greater in the reserve than in the River Great Stour (Harlow 1993; Wain 1999). In winter 2000 and 2001, river input seemed to be a significant feature of the water balance, almost to the same magnitude as stream inflow. However it is impossible to know the importance of its influence in drier years. Again, using the alternative water balance analysis meant that river input was not included in the final model.

The alternative water balance model used was more conceptual than the original water balance, based on known inputs and the site water requirements. It placed more emphasis on knowing site requirements and whether these could be met, rather than modelling the exact amount of water stored on the site each year. It assumes that outflow is a function of the current level of storage on the site as outflow occurs when water levels reach a threshold. The Lampen Stream flows onto the site and interacts with the ditches that criss-cross the reedbed and wet grasslands, and from these ditches water flows onto the reedbed itself. The water flows along hydrological gradients so the amount of water that leaves the stream depends on the ditch water levels and the height of the site water table. All water in the ditches that has not been lost through evapotranspiration eventually flows back into the main stream channel and leaves the site. The amount of water leaving the site can be controlled by the site manager using a sluice on the Lampen Stream and two pen stop dams. If more water is required, water can be held on the site. In addition there has been consideration given to installing a wind pump which would allow additional lifting of water from the Lampen Stream.

The model was created on an annual basis. A year was run from April to March. In terms of storage on the site, each year was made independent of antecedent conditions by setting storage volume on the site on the 1st April at 500 000 m^3 (206 mm) which is around the ideal level for wildlife on the site at this time of year, particularly Bitterns. This was justified conceptually by imagining that either the water availability from rainfall was sufficient to meet this need or that streamflow and abstraction were able to supply any deficit. However the model is not entirely independent of antecedent conditions as the rainfall-runoff models are run continuously throughout the data time-series. Therefore the stream discharge on 1st April depends on the previous year's conditions. This means that if the 27 years considered were arranged in a different order, slightly different results would be expected. In reality conditions in any one year are related to the conditions of previous years and so there are an almost infinite number of combinations of present and past climate. A thirty year time series captures a certain amount of this variation but cannot hope to include it all.

Initially the importance of the water from the Lampen Stream to the reserve was investigated using the rainfall – evapotranspiration balance. On an annual basis it was found that in 90% of years investigated there was a positive balance and the reserve would be able to sustain itself with no addition of water from the Lampen Stream so this indicates that the reserve appears to be sustainable as a wetland in the long term. However it is interesting to note that all three years with a deficit occurred in the 1990s (1990, 1995, 1996 – the years that were the impetus for this research) and this could indicate a possible increase in the frequency of water deficient years. The possible impact of climate change is investigated further in Chapter 7.

For *Phragmites* and many reedbed birds, it is important that there is an acceptable amount of water available all year round. In each year (except 2000 which has the highest summer rainfall of all the years studied and 1987 which has the second highest summer rainfall and the third lowest total of evapotranspiration), the balance of evapotranspiration and rainfall leads to a summer deficit and potentially water falling below ideal levels. The model was based on the principle that all the discharge at the inflow point on the Lampen Stream would flow through and leave the site unless the mean water depth was below a threshold, set at 100 mm based on Bittern requirements. If depth were below this threshold water would flow onto the site from the Lampen Stream until the threshold was reached. It was also assumed that when levels were above an upper threshold water would leave the site as outflow. The total amount of water required from the Lampen Stream to keep the mean water depth on the site above 100 mm was calculated for each summer. In nine out of twenty nine years no water was required. Rainfall alone was enough to maintain the site. In reality water may or may not come onto the site from the stream to take water levels up towards the upper threshold. Overall there is a 60% chance that less than 10 000 m^3 will be required from the stream (including 31% of years when none is required). There is a 20% chance that more than 20 000 m^3 will be required and the maximum is nearly 50 000 m^3. It is important that this is related to the water resource that is actually available because it is likely that in the years with the most need, the low rainfall and high evapotranspiration causing the deficit will also reduce stream discharge. In 60% of years investigated less than 20% of the total discharge of the Lampen Stream at the inflow point of the site will be required to maintain the site and in 78% of cases less than 50% will be required. However in around 14% of years, even the entire discharge of the Lampen Stream for the whole summer will not be enough to meet site water requirements. In these cases, either water must be taken from elsewhere, or less than perfect conditions must be accepted in these years.

In reality, however, there are limits on how much water English Nature may remove from the Lampen Stream. Other users have rights to the water, for example neighbouring farmers have abstraction licences for irrigation purposes and thus have first call on the water (Burnham 1999). The other abstraction licence holders upstream of Stodmarsh require between 5 and 27% of the inflow. When conditions are driest and demands from the reserve highest, demands from the abstractors will also be highest, and the water resources scarcest, leading to potential conflicts over water use. When abstraction is included, 5 out of 29 years (17%) have water demands greater than the entire summer discharge of the Lampen Stream. The proportion of years requiring more than 50% of the Lampen Stream increase from 22% to 36%, and more than 58% of years require more than 25% of the Lampen Stream. Even after abstraction needs have been accounted for it is unlikely the stream would be allowed to dry out. The stream is a habitat in itself and flows into the River Little Stour, which is a Biodiversity Action Plan priority habitat as it is a chalk river. It has also been identified as being potentially impacted by abstraction, hence the setting up of the Little Stour Alleviation of Low Flows project (Environment Agency 2002). These considerations may also limit water availability for use on the reserve.

6.6 CONCLUSIONS

Inadequate modelling of the stream outflow from the reserve led to the use of an alternative water balance model to that defined in Chapter 5, more focussed on assessing adequacy of water availability than accurately modelling storage levels throughout each year. It was found that in 90% of years, on an annual basis rainfall was enough to maintain site water levels. In almost all years however there is a deficit in the rainfall – evapotranspiration balance in the summer. It was found that in two thirds of summers, water is required from the Lampen Stream to prevent the mean water depths on the site falling below acceptable levels for wildlife. In most years less than 50% of stream water is required but in 0.14 years 100% of the water would be insufficient and the site would be drier than would be ideal. This proportion increased when the water requirements of other abstraction licence holders were taken into account.

CHAPTER 7 – The Impact of Climate Change on the Hydrology of Stodmarsh National Nature Reserve

7.1 INTRODUCTION

Conventional water resources assessments assume that past climate is the key to predicting future conditions. However in the last century there has been climate change of a scale and speed not seen since records began. A thirty year sequence of past weather, the conventional period established by the World Meteorological Office to define climate, is therefore no longer considered sufficient to define the probabilities of certain weather extremes occurring in the future (Hulme *et al.* 2002).

The issue of the hydrology and water requirements of Stodmarsh National Nature Reserve (NNR) was first brought to the attention of the site managers by a series of droughts in the 1990s. The River Great Stour has seen a succession of low flows with droughts 1988-96 and a trend of decreased flows 1961-90 (Arnell 2002). 1988-92 was the worst multiyear drought this century and ground water levels did not approach normal in parts of Kent until winter 1993-94 (Holt and Jones 1996). From 1995, in response to these droughts, the government began a programme to co-ordinate long-range water planning in England and Wales, considering changes in supply and demand due to climatic and non-climatic trends. The government asked the private water sector to prepare detailed plans for adapting to future global warming in the UK (Subak 2000). Climate change is of concern to conservationists as changes in the physical environment potentially occur faster than flora and fauna can adapt to them (Agnew and Fennessey 2001).

7.1.1 Climate change

The climate parameter with most clearly discernible change is global surface temperature. Since the beginning of the twentieth century the global average temperature has risen by 0.6°C, with 0.4°C of the rise occurring since the 1970s (Hulme *et al.* 2002). Nine of the ten warmest years since 1860 have occurred between 1980 and

1997 (Arnell and Reynard 2000). This change in temperature has been attributed to increasing concentrations of so-called greenhouse gases in the atmosphere. These gases – including water vapour, CO_2 and methane – occur naturally, trapping radiation in the lower atmosphere. In increased concentrations however, they result in increased global temperatures. Although clear trends in mean global temperature have been seen, climate variability also occurs naturally from causes such as El Nino and the North Atlantic Oscillation, as well as changes in sunspot number and the Milankovitch cycle.

7.1.2 Impacts of climate change on wetlands

The predicted responses of environments and ecosystems to climate change are many and varied. The effects of climate change on wetland ecosystems are potentially severe but are hard to assess and have been the focus of many workshops, conferences and publications in recent years (Winter 2000). As wetlands are sinks for greenhouse gasses, they are involved in feedback responses and are a key ecosystem when considering the impact of global warming. However there are relatively few studies available to assess the impact of climate change on wetland ecosystems (Gitay *et al.* 2001, Arnell 1996). Although it is now feasible to provide scenarios of potential hydrological changes there are few ecological models which can simulate the effects of change (Arnell 1996). Predicting wetland response to climate change is limited by our understanding of how wetland flora and fauna respond to changes in temperature, precipitation, water levels, water quality and atmospheric carbon concentrations (Burkett and Kusler 2000).

The key hydrological variables for wetlands are water level and the frequency and duration of inundation, both of which would be altered by global warming and this could lead to a change in species composition (Arnell 1996). Climate change will affect the hydrology of individual wetlands mainly through changes in precipitation and evapotranspiration regimes. Small changes in these variables which alter water levels by only a few centimetres will be enough to reduce or expand many wetlands in size, convert wetlands to dry land or shift from one wetland type to another (Burkett and Kusler 2000). As wetlands exist in the transition zone between aquatic and terrestrial environments, they are vulnerable to changes in surface and groundwater hydrology that are beyond the limits of tolerance and adaptation of most wetland species. Fragmentation of wetland habitats means that species cannot easily migrate in response to temperature or water level changes (Burkett and Kusler 2000). In general, climatic warming is expected to start a drying trend in wetlands, though this may largely be indirect due to a change in water level (Gitay *et al.* 2001). Wetlands that depend on precipitation for their source of water are the most vulnerable to climate change. Wetlands that are mainly groundwater fed are least vulnerable because of the buffering of regional groundwater systems (Winter 2000).

A detailed study of the impact of climate change on a wetland is that of Poinani and Johnson (1991) on the North American prairie wetlands where warmer, drier conditions are predicted. A number of potential changes to the wetland are speculated upon – a decrease in water depth, migration of wildlife to different geographic areas and an increase in salinity from lower water levels and more frequent drying. There may be a loss of waterfowl habitats due to increasing vegetation cover, reducing open water areas. A climate based simulation model of the climate cycle of a semi-permanent prairie wetland supported this conclusion.

7.1.3 Simulating future climate change

Future climate change is usually simulated using global circulation models (GCMs). GCMs simulate the dynamic behaviour of short-term weather patterns. They do not produce forecasts of future climate but potential scenarios. The simplest GCM based climate scenario directly takes the change in precipitation, temperature or runoff as simulated by a GCM to indicate what will happen in the catchment of interest. However, this is not appropriate as the GCM may not simulate the climate of the study region well and the spatial scale of the model is usually too large. Instead the GCM can be used to define change and perturb a historical data set, assuming that the estimates of change are reliable, even if the simulated climate is not.

There are many uncertainties in the definition of climate change scenarios. There are a number of scenarios available, each giving quite different results. Emissions scenarios simulating future greenhouse gas release are based on the rate of population growth and economic development, which are inherently uncertain. The earliest GCM models modelled the impact of CO_2 stabilising at twice the current concentration, known as an equilibrium scenario. No dates can be given to these simulations as CO_2 concentration will not stabilise at a given level and the rate at which CO_2 increases depends on emissions scenarios. More recent experiments have simulated a gradual increase in gas concentration – known as "transient" models. These are more realistic but there are difficulties in creating scenarios from these simulations, as changes need to be defined relative to a control simulation with stable CO_2. The transient model may in fact not be any better than the best equilibrium estimate (Holt and Jones 1996).

Emissions scenarios are converted into greenhouse gas concentrations, which is complicated by the fact that chemical transformations in the atmosphere and the relative magnitude of gas sources and sinks are not fully understood (Arnell 2002). Gas concentrations must be converted into climate changes. In order to simulate the response of hydrology to climate change the data then have to be put into a format suitable for use in a hydrological model. This involves the perturbation of existing catchment scale meteorological data which is then used to run a hydrological model with both the perturbed and unperturbed inputs (Arnell and Reynard 2000). Climate change impact assessments then compare conditions that might be expected with climate change to the baseline data.

GCMs operate on a short time scale but large spatial resolution. Therefore the data must be downscaled from the large scale of the climate model to the catchment scale. Scaling down can be done using simple interpolation, a correlation between regional and point weather based on observed weather or the use of nested models at a higher spatial resolution. However given the uncertainties in the model itself, excessive complexity in downscaling is probably unjustified and almost all published studies use simple interpolation (Arnell and Reynard 2000). An additional uncertainty is in defining the input data at an appropriate temporal scale for hydrological analysis. Most scenarios define changes in monthly climate and these must then be translated into daily data. The simplest approach is to perturb a historical daily time series, though this has the problem of preserving the temporal structure. A more complex approach is to use a stochastic weather generator given the information in the climate change scenario. However it is very difficult to develop a simple stochastic model that can simulate accurately the observed temporal structure of UK rainfall. Current climate change scenarios tend to define changes in average climate though hydrologists are also interested in changes in variability, such as the changes in rainfall intensity and the frequency of floods and drought (Arnell and Reynard 2000). However changes in variability are not reproduced as well by models as changes in mean climate (Arnell 2002).

7.1.4 The impacts of climate change on the water balance

7.1.4.1 Precipitation

Precipitation is the main driver of the water balance and so change in its frequency or quantity has big implications for hydrology and water resources and will affect flood and drought frequency. Although there have been clear recent trends in global temperature data, the resulting trends for precipitation are less clear. There does appear to have been a shift in temporal and spatial rainfall patterns however, with a general increase in precipitation in the Northern hemisphere (Arnell and Liu 2001) and a tendency towards wetter winters and drier summers (Arnell and Reynard 2000). Osborne *et al.* (2000) studied 110 sets of rainfall data from between 1961-1995 and found that in winter there had been a shift away from medium rainfall events to a greater contribution from heavy daily amounts at most locations across the UK. In summer there was a decreased importance of heavy events and more lighter and moderate events. The frequency of extreme rainfall events may also have increased. Recent scenarios for the UK indicate a future increase in the variability of seasonal and annual rainfall and more heavy rainfall events (Arnell and Liu 2001).

7.1.4.2 Evapotranspiration

Increasing temperature may result in an increase in evapotranspiration because the water holding capacity of the air is increased. Budyko (1982) estimated that potential evapotranspiration would increase by 4% for every 1°C temperature rise. However evapotranspiration is influenced by many factors which may be affected by global warming which may exaggerate or offset this rise. Net radiation may be reduced by increased cloudiness. Incoming long wave radiation is affected by atmospheric conditions including humidity levels and CO_2 concentration, whilst outgoing long wave radiation is affected by surface temperature. Reduced radiation may lead to reduced evapotranspiration demands and this will have a greater effect in humid regions such as the UK as atmospheric moisture content is a major limitation to evapotranspiration. Transpiration may also be affected by changes in plant characteristics. Increased CO_2 concentration may increase plant water use efficiency, reducing stomatal conductance and therefore reducing transpiration in some species, though this may be counteracted by increased plant growth (Arnell and Liu 2001). However these plant factors are based on experiments under controlled conditions and it is hard to be sure as to what will happen in reality and plants may adapt to increased CO_2.

Due to this complicated range of factors affecting evapotranspiration there is no consensus as to even whether it will increase or decrease (Arnell 1996). Arnell and Reynard (1996) prepared two scenarios based on a sensitivity analysis of the Penman Monteith equation. In their first scenario only a change in temperature was assumed, and a second scenario included changes in net radiation, windspeed and relative humidity. Little information was found as to the direction or magnitude of potential change in relative humidity. It could be increased by the fact that increased temperature increases evapotranspiration and therefore the water vapour content of the air. Alternatively the fact that a temperature increase increases the deficit between specific humidity and the saturation of the air, increasing the potential for further evaporative losses could cause a decrease. This second factor seemed dominant so relative humidity was assumed to decrease by 7% throughout the year. The first scenario resulted in an increase of evapotranspiration by 9-11%, whilst the second results in an annual increase of 29-36%. Using Penman Monteith it was found that evapotranspiration was most sensitive to changes in temperature, net radiation and stomatal conductance.

7.1.4.3 Streamflow

A number of studies have been carried out on the impact of climate change on runoff and streamflow in the UK. In general changes identified are consistent with patterns of change for rainfall and generally runoff increases where rainfall increases. Seasonal patterns may be exaggerated but the overall timing of flows will be unaffected. As a general rule, the lower the current ratio of runoff to precipitation, the greater the relative effect of a given change in precipitation on streamflow (Arnell 2002). However it is hard to see long term trends in historical hydrological data as records are generally short and most catchments have been exposed to human intervention.

A study investigating the impact of climate change on British rivers has been carried out using data from a large number of catchments (Arnell 1996; Arnell and Reynard 1996; Arnell and Reynard 2000). Both an equilibrium scenario representing conditions in 2050 and transient scenarios between 1990 and 2050 were used. A lumped conceptual model was used to simulate runoff and it was assumed that all catchment properties remained the same. They found that the effects of climate change varied seasonally. At the River Medway in Kent, under the driest scenario there was a decrease in runoff all year round but under the wetter scenarios there was an increase in winter and a decrease in summer. Generally there was an indication of a reduction in streamflow in Southern and Eastern UK and an increase in the North. The greatest percentage change in runoff occurred in catchments with the lowest proportion of rainfall going to runoff. Nationally there was a tendency to increased seasonality of flow, with more runoff in winter and a reduction in summer. There was a large difference between the different climate change scenarios. The wettest scenario would imply an increase in average runoff whilst the driest would result in a reduction of up to 30%. The drier areas of the South and East show the greatest sensitivity to climate change. A change in the number of days when rain falls has a smaller effect on runoff than changes in mean monthly rainfall.

Sefton and Boorman (1997) also studied possible impacts of climate change on UK streamflow. IHACRES was run using historical and perturbed historical data. This was extended to ungauged catchments by estimating model parameters from relationships between parameters and physical attributes of the catchment. The UK Hadley Centre's high resolution (UKHI) model was used. Reductions in flow were found in the centre and East of the country, and there were increases in the North and West. However flow changes were not always correlated with climate but were also influenced by the nature of the catchment – steeper upland catchments show the greatest increase in discharge whereas reductions in flow corresponded to the permeable geologies of central England.

Holt and Jones (1996) studied water resources in Wales using UKHI and UKTR (the transient model). UKHI predicted that even under the wettest scenario there could be reduced resources in summer and autumn due to increased evapotranspiration. The best estimate gave a 20% reduction in discharge. UKTR produced predictions that fell between the wettest and best UKHI scenarios. Predicted flows are above present levels in winter and spring, whereas in summer and autumn higher temperatures and reduced precipitation meant a reduction in runoff.

Pilling and Jones (1999) adopted a higher resolution model than previous studies in considering the potential effects of climate change on British runoff using the Hadley centre's high resolution (UKHI) and transient (UKTR) models, interpolated onto a 100 km^2 grid. Both models predict a 5-10% increase in mean annual precipitation with an increase in winter precipitation of 10-25%, and a 0-10% summer decrease in South-East England. Under UKHI this results in predictions of increased mean annual effective runoff over most of Britain though there are decreases in the South-East. There is an increase in winter runoff over the whole of the UK except in South-East England. In summer there are reductions of greater than 30% around the Thames estuary and more than 70% of Britain experiences some reduction in summer runoff. In general runoff increases in the North while decreasing in the South-East and there is increased seasonality in the water balance.

All studies using hydrological models assume that only the inputs to the catchment will change, when in reality climate change will also change processes operating within the catchment. For example, soil properties may be affected. Higher temperature and an increase in precipitation would lead to a loss of soil organic matter and a decrease in the ability of soils to hold moisture. Increased desiccation may enhance soil cracking and lead to an increase in infiltration. Increased precipitation may increase gleying, reducing infiltration.

7.1.5 Climate change scenarios for the UK

Climate change scenarios have been developed for the UK by the UK Climate Impacts Programme (UKCIP) and the most recent set of scenarios is referred to as UKCIP02 (Hulme *et al.* 2002). UKCIP02 uses four potential emissions scenarios (Table 7.01), based on a report by the Intergovernmental Panel on Climate Change (IPCC) (IPCC 2000). It is not possible to assign probabilities to these scenarios as which, if any, occurs in reality depends on many factors such as population growth, economic growth and global warming mitigation strategies.

Table 7.01 – The four emissions scenarios used by UKCIP02

UKCIP02	SRES*	2020s		2050s	
emissions scenario	scenario	ΔT(°C)	CO_2 (ppm)	ΔT(°C)	CO_2 (ppm)
Low	B1	0.79	422	1.41	489
Medium-low	B2	0.88	422	1.64	489
Medium-high	A2	0.88	435	1.87	551
High	A1FI	0.94	437	2.24	593

* IPCC Special Report on Emissions Scenarios (IPCC 2000)

For each emissions scenario, climate change scenarios were created based on experiments using the HadCM3, HadAM3H and HadRM3 climate models which are based in the UK and are among the most advanced in the world. HadCM3 is the latest fully coupled atmosphere-ocean GCM which produces patterns of climate change across the whole earth. Its output is used to drive a higher resolution (~120 km) model of the global atmosphere (HadAM3H). This then is fed into the high-resolution model of the European atmosphere (HadRM3).

The climate change scenarios are based on a comparison with 1961-1990 baseline data and the climate change scenarios are presented as a map showing changes for three thirty year periods, the 2020s (2011-2040), the 2050s (2041-2070) and the 2080s (2071-2100).

For each scenario, the UK is represented by a grid of 104 boxes that give a resolution of around 50 km. 50 km changes in average climate are combined with 5 km observed climate data set to produce descriptions of future annual climate. Temperature and precipitation data are available but evapotranspiration data is not provided and must be calculated. For all variables except precipitation, the data is based on simple interpolation of the 50 km grid onto a 5 km grid used for the baseline historical meteorological observations. For precipitation data, however, percentage rather than actual values are used in the downscaling process, reflecting local topographical variations.

7.2 AIMS AND OBJECTIVES

The aim of this section is to use the models of the hydrology of Stodmarsh National Nature Reserve created in Chapter 6 to make predictions of the likely impacts on the hydrology of the site of climate change, as predicted by the UKCIP climate change scenarios on the 5 km resolution using the medium-high emissions scenario.

7.3 METHODS

Data from UKCIP 2002 (downloaded with permission from http://www.ukcip.org.uk/scenarios/index.html) on the 5 km grid scale was used to create climate change scenarios for Stodmarsh. Scenarios were created for the medium high emissions scenario, which is characterised by self-reliance, the preservation of local identities, a continuously increasing population and economic growth on regional scales. Scenarios were created for the 2020s and the 2050s.

Data for rainfall and temperature were available on a 5 km scale grid. Evapotranspiration had been previously estimated from available data using the Penman Monteith equation. The UKCIP 5 km data set contains predicted monthly averages of maximum and minimum temperature and windspeed. In order to estimate reference evapotranspiration, solar radiation and relative humidity are also required. Monthly mean solar radiation was estimated from latitude, date and temperature using the method of (Hargreaves et al. 1985) calibrated against the baseline (1961 – 1990) data. Similarly, mean relative humidity was estimated by assuming dew point temperature equal to minimum air temperature.

The grid square including Stodmarsh NNR was chosen. UKCIP provide mean monthly data for each parameter from their baseline data for the period 1960-1990 and predicted mean values for the 2020s and 2050s. Rather than using the future climates for the 2020s and 2050s as predicted by UKCIP, it is more accurate to find the change from current ambient conditions, due to possible discrepancies between their baseline data and that calculated for Stodmarsh. The absolute change between the UKCIP baseline and 2020 and 2050 was found and this was applied to a baseline created from the meteorological data previously used in this study. Historical data for Stodmarsh was unavailable for the whole of the period 1960-90 and therefore the following periods were used:

Precipitation: 1964-1990

Temperature: 1964-1990

Evapotranspiration: 1973-1990

There is debate over the merits of using absolute change as opposed to the proportional change. The absolute change is generally considered to be more accurate, as with proportional change, if the baseline data is very different, or if extreme weather days occur, this can lead to unrealistic magnitudes of change. It will also lead to greater impacts of climate change in wet and warm years than in dry and cold years which are not necessarily realistic.

In the case of temperature and evapotranspiration, for each month of historical data the absolute mean monthly change for that month was divided by the number of days in the month and then this change was added to the existing datum for each day. In the case of rainfall the absolute mean monthly change in rainfall was divided by the number of days with rainfall that month, and applied to each rain day. Where the mean rainfall decreases, this could lead to apparent negative rainfall on days where rainfall total is small. Where this occurred, the value for that day was set to zero, and the amount by which it had become negative subtracted from the next sufficiently large rain day that was within the same month. In this way the total rainfall change in a month is kept consistent with the model. Although no conscious effort was made to alter the frequency of rain days in a month, this approach meant that in months where total rainfall was reduced, the number of days with rainfall was in some cases also reduced.

The resultant evapotranspiration and rainfall data were fed into the hydrological model for Stodmarsh described in Chapter 6. The data was also fed into the rainfall runoff models in order to create simulated future runoff scenarios which could also be used in the overall hydrological model.

7.4 RESULTS

Figure 7.01 shows mean monthly precipitation, temperature and evapotranspiration for the 5 km grid square that includes Stodmarsh, for the historical baseline data and the 2020s and 2050s.

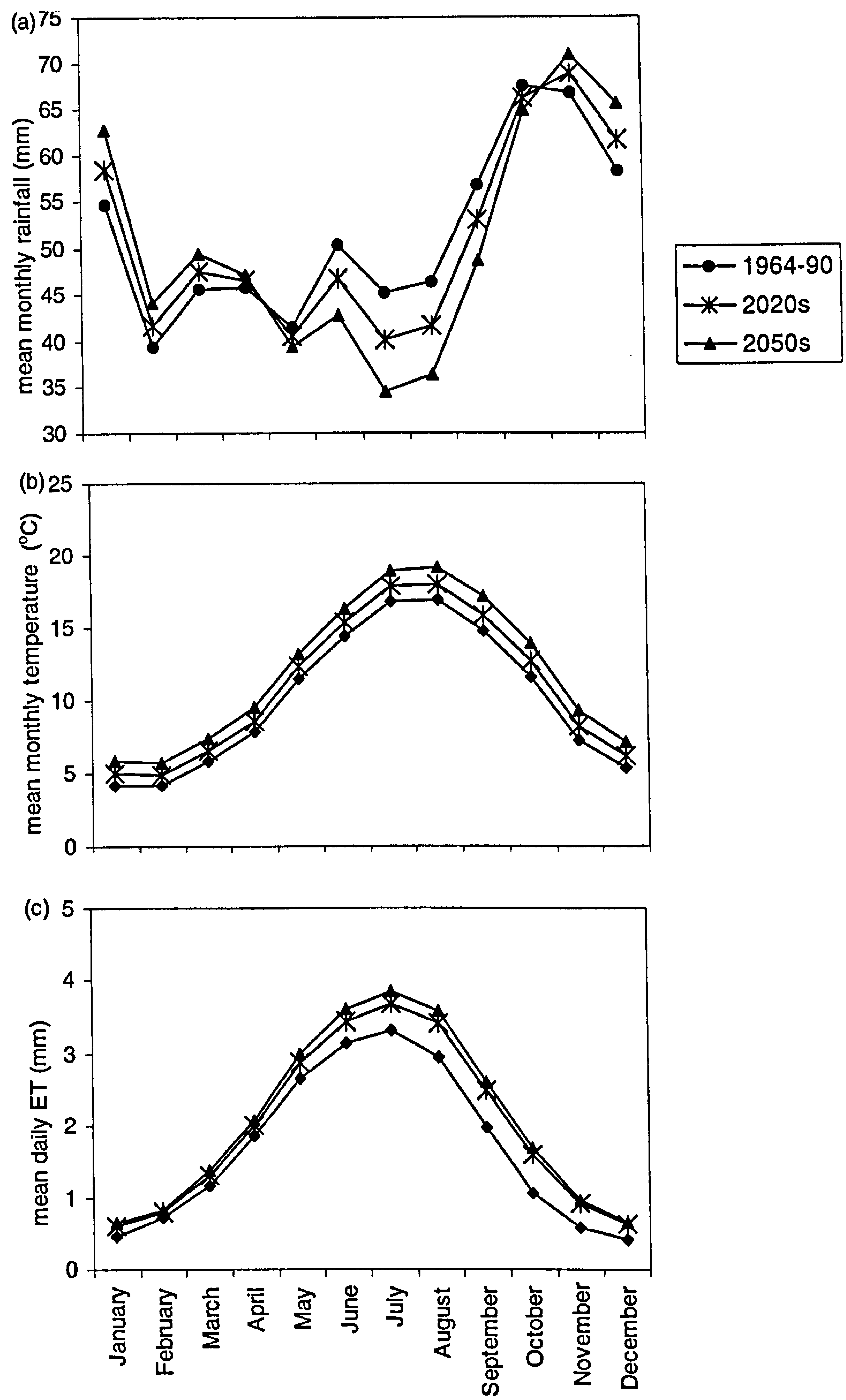


Figure 7.01 – Current and predicted monthly mean (a) precipitation, (b) temperature and (c) evapotranspiration for the 2020s and 2050s under the medium-high scenario for Stodmarsh National Nature Reserve.

Between November and April there is a predicted increase in precipitation of between 0.6 mm and 3.9 mm per month in the 2020s and between 1.4 mm and 8.2 mm per month in the 2050s compared with the baseline data. From May to October precipitation is predicted to reduce between 0.9 mm and 5.0 mm in the 2020s and 2.0 mm and 10.7 mm in the 2050s. On an annual basis these two effects go some way to cancelling one another out so that the net annual effect is a reduction by 5.3 mm by the 2020s and by 11.3 mm by the 2050s. In all months of the year the mean temperature is increased by between 0.7°C and 1.1°C in the 2020s and between 1.6°C and 2.4°C in the 2050s. Evapotranspiration is also predicted to increase throughout the year. The increase is much greater from June to December than from January to May.

The change in the rainfall–evapotranspiration balance in the summer (April to September) and winter (October to March) on Stodmarsh NNR was investigated under the climate change scenarios. Under climate change the difference between summer and winter is increased. In almost every year the water deficit in summer becomes greater, as does the surplus in winter. This means that on an annual basis there is no change in the number of years that have a negative rainfall-evapotranspiration balance under climate change as compared to current conditions. However it would be expected that the increased summer deficit would lead to a greater demand for water from the Lampen Stream or other sources in order to maintain the site in optimum conditions. The same water requirement model was used as is described in Chapter 6. This is a water balance based on rainfall inputs and evapotranspiration losses. The aim of the model is to work out how much water is required from the Lampen Stream in order to keep water levels on the site between the ideal depths of 100 mm and 250 mm. It is assumed that when the mean site depth is above 250 mm water is drained from the reserve. When levels drop below 100 mm, water comes onto the site from the Lampen Stream either through natural gravity flow or through management. The total volume of water required from the Lampen Stream can be calculated for each summer in the twenty nine year historical data series under each climate change scenario. This can then be plotted as a probability (Figure 7.02).

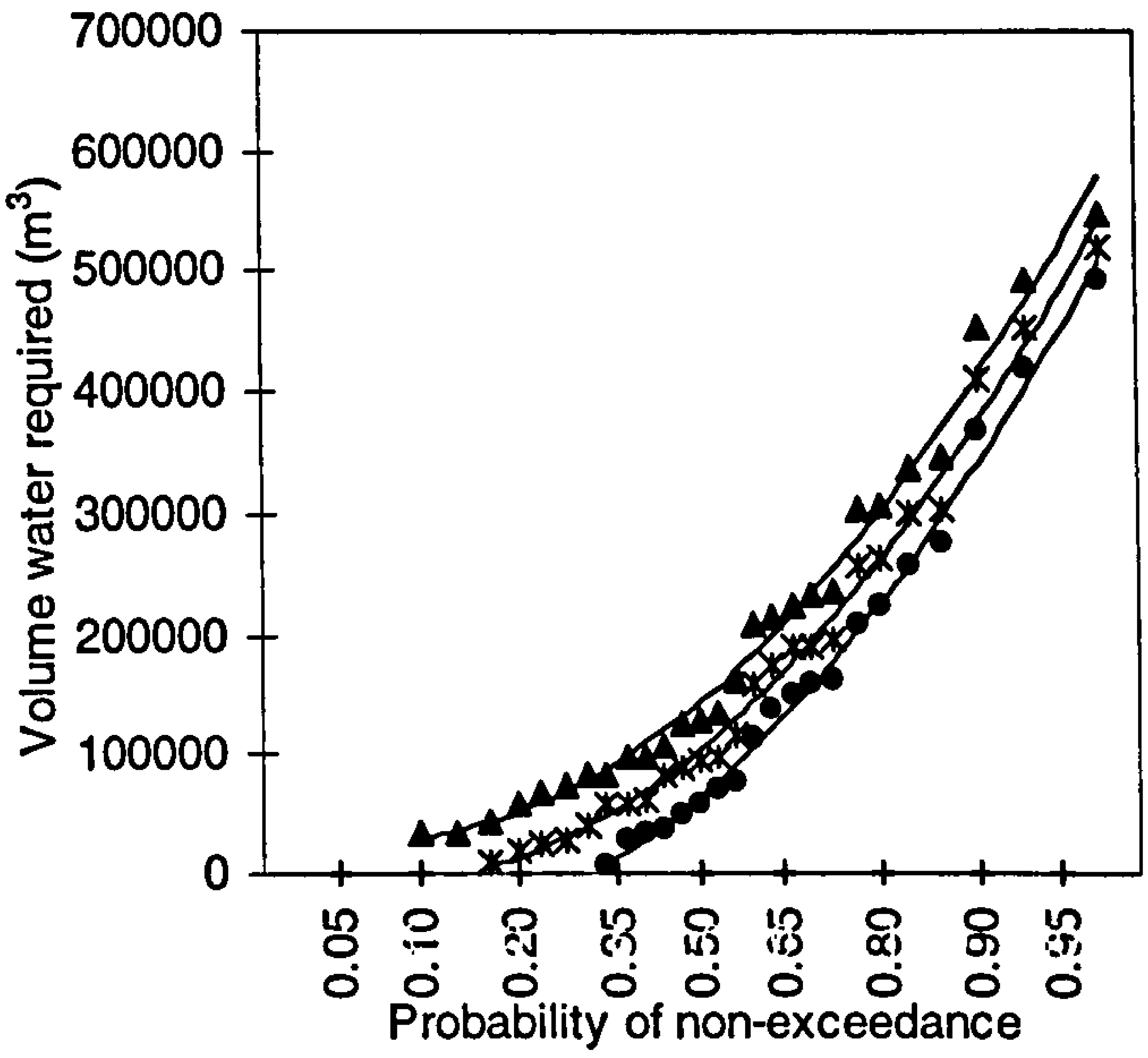


Figure 7.02 – Probability of requiring different volumes of water during the summer 1974-2001 (circles), 2020s (stars) and 2050s (triangles).

An important change under the scenarios of future climate change is in the number of years that require no water at all. Under historical conditions this was 9 years out of 29 (0.31 chance). In 2020 this had dropped to 4 years (0.13) and under 2050s conditions to 2 years out of 29 (0.07). As expected, climate change increases the chance of requiring higher volumes of water. Under current conditions. in 60% of years there is likely to be a requirement for less than 10 000m^3 water. However in 2020 this will occur in only 56% of years and in 2050 in only 42%.

There is also a change in water availability due to change in stream discharge. The change in precipitation was fed into the IHACRES inflow model and the mean monthly total inflow discharge for the Lampen Stream can be seen in Figure 7.03.

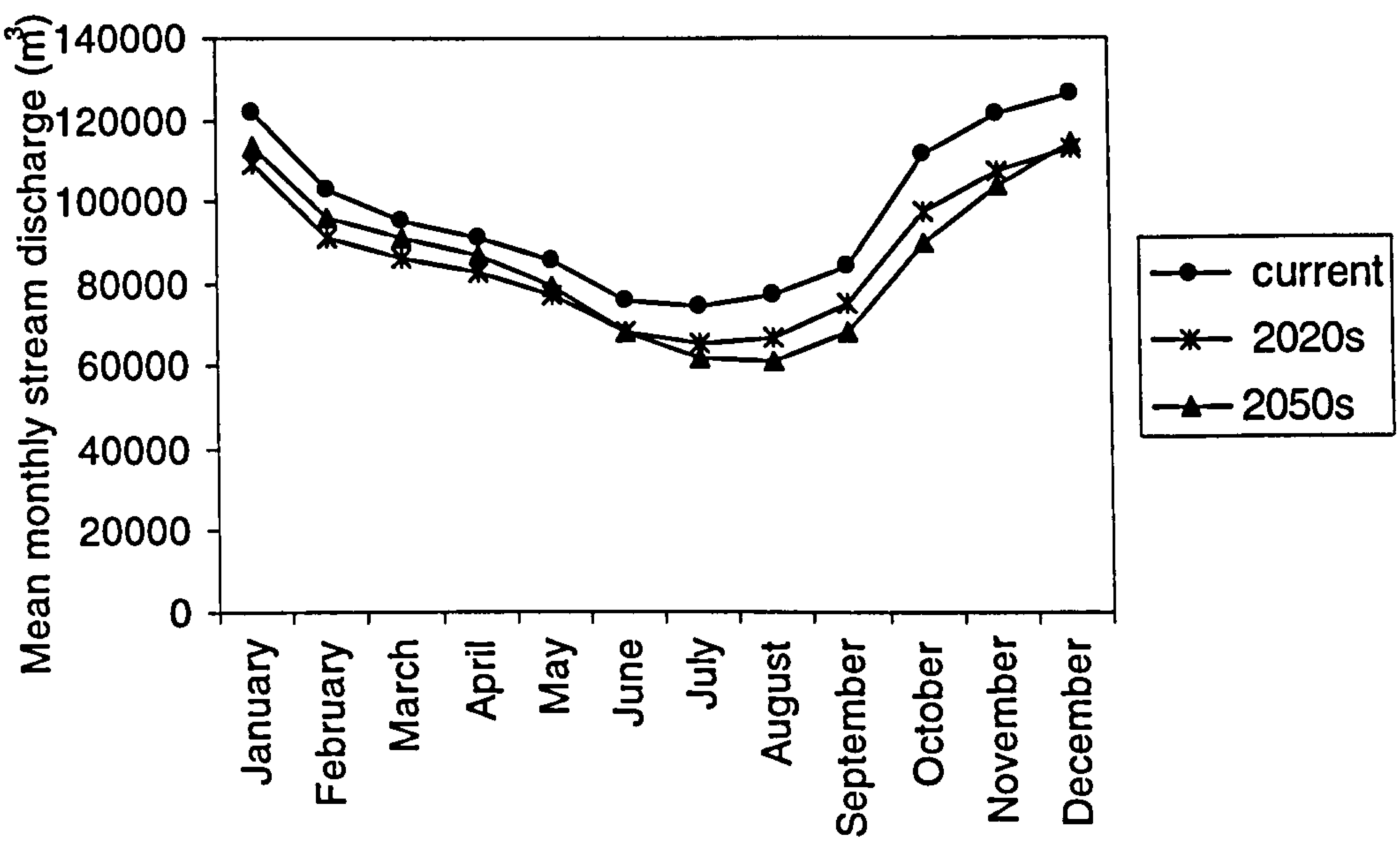


Figure 7.03 – Current and predicted monthly mean total flow in the Lampen Stream for the 2020s and 2050s under the medium-high scenario.

It can be seen that climate change results in a reduction in inflow at all times of year for both the 2020s and 2050s compared to current conditions. Between December and May, the reduction is greater in the 2020s than in the 2050s. The reduction between current conditions and the 2020s is greater than that between the 2020s and the 2050s.

The volume of water required was calculated as a percentage of water available in the Lampen Stream as determined by the IHACRES inflow model (Figure 7.04).

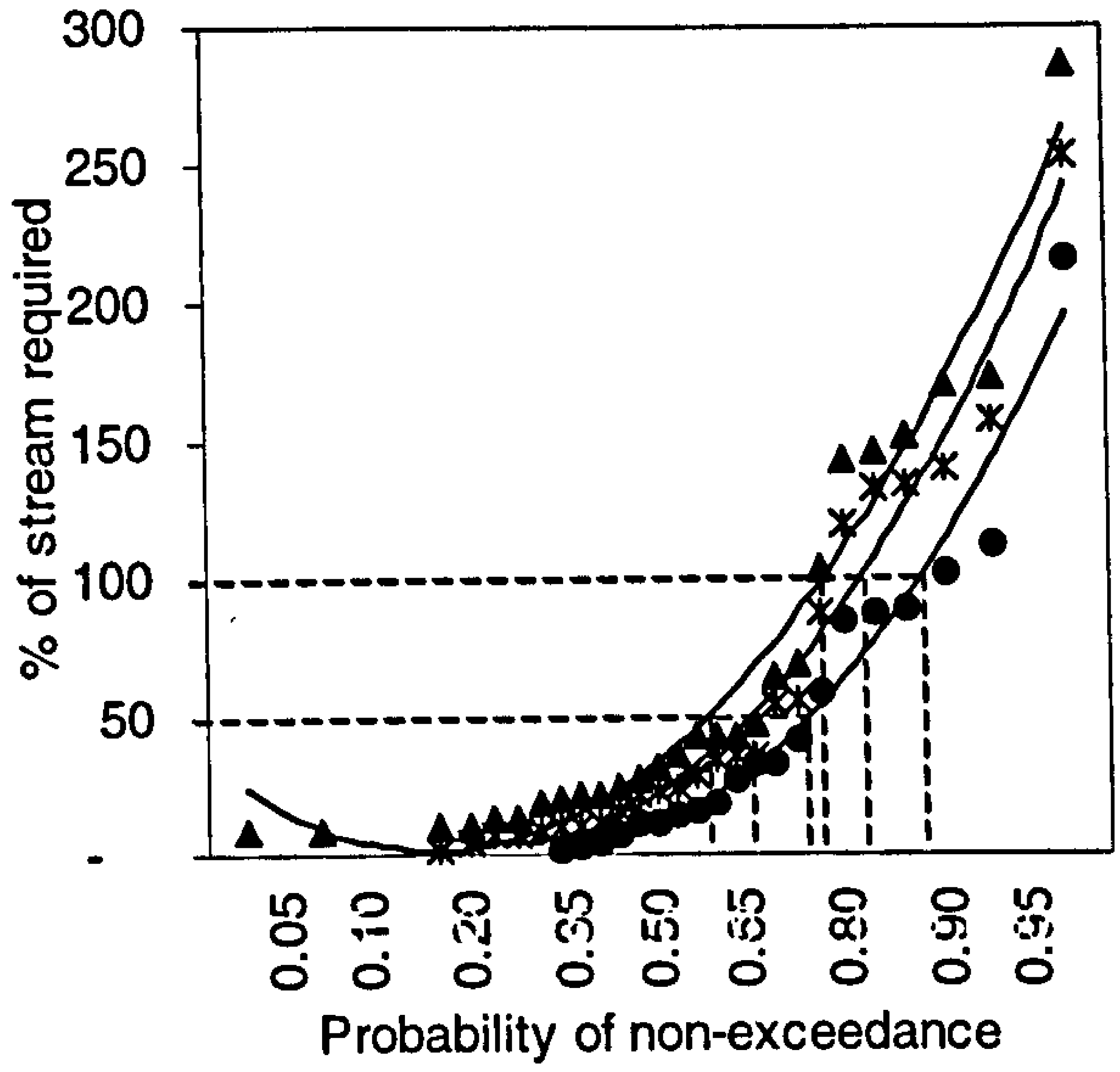


Figure 7.04 – The probability of requiring various percentages of the inflow as determined by current conditions (circles), 2020s (stars) and 2050s (triangles). The dashed lines show the probability of requiring more than 50% and 100% of available Lampen Stream water.

Climate change will lead to demand for an increased proportion of Lampen Stream water on the reserve. This is summarised in Table 7.02.

Table 7.02 – The probability of Stodmarsh NNR summer water requirements exceeding 25%, 50% and 100% of inflow discharge in the Lampen Stream.

	> 25% of inflow	> 50% of inflow	> 100% of inflow
Current	0.36	0.27	0.14
2020s	0.46	0.36	0.22
2050s	0.55	0.44	0.29

The table shows that requirements exceed the total discharge of the Lampen Stream more than twice as frequently in the 2050s as with current data. At every level there is a significant increase in water requirements in the summer under climate change.

7.5 DISCUSSION

Any climate change impacts assessment includes many uncertainties at every stage of the modelling procedure. Different models give different results of the impact of global warming. In the above analysis only one climate change scenario is used – "medium – high" which is the scenario that represents something close to "business as usual" and assumes no major mitigation of greenhouse gas emissions. The UKCIP include four scenarios and in reality there are an infinite number of possibilities. Climate change may be reduced if governments begin to seriously tackle levels of gas emissions, but equally population and economic growth and industrialisation, particularly in developing nations, may lead to greater emissions and greater global warming. Incomplete understanding of atmospheric conditions and chemistry led to uncertainty in the impact of these gases. This uncertainty in the climate change scenario is added to the uncertainty in the hydrological model (considered further in Chapter 8) with the net consequence that all results must be taken simply as possibilities and not as predictions of the future.

In addition a number of assumptions are made in the above analysis which may be unrealistic. The first is that catchment conditions will remain the same and the parameters used in the models will remain unchanged. In reality there may be changes to soil structure, land use and vegetation cover, which would change the mechanisms of hydrological responses. The second assumption is inherent in the methodology used, namely in perturbing a historical data set. Although this is the easiest and most commonly used methodology it has the problem of preserving the temporal structure of the weather. Climate change modellers have predicted a change in the frequency of extreme events and this is not included in the above analysis. UKCIP02 does not give clear indications of the changes in the frequency of rainfall events that can be used to perturb historical rainfall data. However changes in timing frequency and magnitude of extreme rainfall events are probably more important to wetlands than changed mean conditions as the former change could alter the annual water level regime (Agnew and Fennessey 2001).

As would be expected from the primary impacts of climate change on the water balance, (increased winter rainfall, decreased summer rainfall, increased evapotranspiration) the hydrological model predicts is an accentuation of the difference in summer and winter conditions. As was noted in Chapter 6, there is a tendency on the site to have a deficit in the rainfall–evapotranspiration balance in the summer and a surplus of water in the winter. These surpluses and deficits will be increased under climate change. The changes appear small in comparison to inter-annual variation but have significant water resource implications. The major period of concern is the summer when there is potential for the site to dry out to an extent that is detrimental to the wildlife on the site (it is much easier to remove winter surpluses from the site than to compensate for summer deficits).

Streamflow in general is decreased, with the greatest decreases occurring in the summer months, coinciding with the greatest reduction in rainfall. The increases in rainfall in winter are not enough to bring streamflow up to above current levels and winter streamflow in both 2020 and 2050 is similar to or less than current levels. This concurs with the work of other authors such as Arnell and Reynard (1996) and Pilling and Jones (1999) who also found a year round reduction in runoff in South-East England. A reduction in resources increases the potential problem of meeting summer deficits.

The water demand model demonstrated that under the medium-high climate change scenarios there would be an increased demand for water from the Lampen Stream both in terms of absolute volume of water and in terms of percentage of the Stream required. Currently nearly one third of summers require no water at all, but this drops to only13% in the 2020s and only 7% in the 2050s. When including resource availability, the most serious situation is when demand cannot be met even by using all the Lampen Stream discharge over the whole summer. Currently this occurs in 14% of years. However in the 2020s this could increase to 22% and in the 2050s to 29%. A site manager may accept conditions below optimal in just over one in ten years. However as this recurrence increases towards one in three years this is likely to prove less acceptable and much more of a threat to the long term sustainability of the site. In this situation it would be necessary to consider abstracting water from another source, or storing water from the winter months. However to be able to take 100% of the Lampen Stream water is an impossibility. Farmers also have abstraction rights for irrigation (and their needs are also likely to increase under climate change). If a slightly more realistic limit of 50% of the Lampen Stream water is set, the proportion of years that remain short of water rises from the current value of 27% to 44% in 2050.

However changes in the frequency of extreme events could make the implications of climate change even more serious. In the UK the incidence of drought is predicted to rise. Summers as hot and dry as 1976 (the year with the greatest water requirements requiring 215% of the Lampen Stream under non-perturbed conditions) have been predicted to increase in frequency one hundred times to one year in ten (DoE 1991). More recent models have shown that hot summers such as 1995 (with the next highest water requirements, needing 89% of the Lampen Stream) will increase from being a one in ninety year event, to being a one in three year event (DoE 1996). Climate change can therefore be said to have serious implications for the water resource availability of the Lampen Stream catchment and Stodmarsh National Nature Reserve and the long term sustainability of the wildlife interests of the area.

7.6 CONCLUSIONS

The uncertainties when modelling using future climate change scenarios are vast. However the models show that in summer months the rainfall – evapotranspiration deficit will increase leading to a greater demand for water from the Lampen Stream. At the same time discharge within the Lampen Stream will decrease, reducing resource availability and increasing the proportion of Lampen Stream discharge that will be required in order to maintain the site. The number of years where even the total of summer discharge would be insufficient will increase from 14% to 29% in 2050, and if abstraction limits were set at 50% of the Lampen Stream, this would result in 44% of years requiring water from an additional source in order that the reserve is maintained in optimal condition. These changes in mean climate may well be exacerbated by changes in frequency in extreme events, leading to further increases in water requirements and the potential for site damage.

CHAPTER 8 – Quantification of Error in the Water Balance

8.1 INTRODUCTION

In order to assess the confidence that can be placed in the water balance that has been created for Stodmarsh National Nature Reserve, it is necessary to make an assessment of the likely errors within it. It is widely acknowledged that accurately characterising wetland hydrology can be difficult. Water budget calculations are often subject to large errors due to the difficulty in measuring the various components (Winter 1981; Arnold *et al.* 2001; Lott and Hunt 2001). Not evaluating these measurement errors means that it is impossible to differentiate between a poor water balance and a good one, which can be very misleading.

When a complete measured water balance is calculated, the error is usually estimated as the difference between the inflows and the outflows (the "closed balance" approach) (e.g. Koerselman 1989; Gilvear *et al.* 1993). This is a useful technique but it does little to ascertain whether the methods used to estimate each component contain an acceptable amount of error. A small value of error for the complete balance is no guarantee of the absence of measurement errors, since errors may be compensating (Drexler *et al.* 1999).

Error analysis is especially important when using residual terms, such as using the water balance to estimate change in storage. Lack of error analysis results in the residual term having little meaning. Assuming errors are random, the more data that is considered, the more errors will compensate and therefore long term estimates have lower error rates.

There are some examples of attempts at rigorous error analysis of wetland water balance calculations. Owen (1995) calculated errors both using a closed balance technique and from the standard deviations of measurements and instrumental error for each component of the hydrologic budget, summed into the final mass balance. Errors of 7.1% and 4.5% were found over the two seasons measured, based on the closed balance. Errors for the overall balance based on the instrument and measurement errors were much larger. However as the water balance almost balances it was assumed to be a close approximation of reality.

Drexler *et al.* (1999) carried out error analysis for each component of the water budget of a 1.5 ha peatland in New York State U.S.A. The errors in precipitation were associated with raingauge placement and instrument error and were approximately 10% at a gauge density of less than one per 21 km^2. Streamflow estimates contain error associated with stream-discharge calibration; model error and stage measurement error were large due to the use of a small weir. MORECS (Meteorological Office Rainfall and Evaporation Calculation System) data was used to estimate evapotranspiration, which resulted in a very large error (40%) as it was calibrated over grassland due to the limited knowledge of bryophyte evapotranspiration. Groundwater errors were also around 40%. They compared their error rates with other authors including those of Owen (1995) and Gehrels and Milamootti (1990) and found the errors to be well within the range of those of other studies. They concluded that there is a wide margin of error for all components except precipitation and that the data demonstrated the difficulties in accurately quantifying hydrologic budgets of wetlands. Better methods of measurement of individual components would lead to improvements in water balance estimation and in the understanding of wetland hydrology.

Gehrels and Milamootti (1990) carried out a twelve month water balance study on a small *Typha* marsh. Three approaches were used in the quantification of errors: closed balance, a chloride balance and theoretical error analysis. The closed balance was done over periods with little storage change and resulted in errors of 4% to –9%. For the whole year the chloride budget estimated errors of around 10%. Hydrological and mass balance error ratings were assigned to each component of the water balance based on statistical techniques and extrapolation of values reported in the literature. The total theoretical error for inflows was 21% and for outflows was 30% which would result in a large error over the whole balance (the sum of the absolute errors).

The Bowen ratio energy balance method was used to estimate a part of the measured water balance created in this study. Because it involves the use of a combination of instruments, a detailed error analysis is potentially complex. Several authors have attempted error analysis of the technique. Revfeim and Jordan (1976) calculated an error in evapotranspiration of 10% which is often quoted but this was calculated using older and less reliable equipment than is used today as vapour pressure was calculated from wet bulb temperature. Fuchs and Tanner (1970) quantified errors using a simple additive approach. They found errors within all the measurements made and that a larger value of the Bowen ratio resulted in larger errors. The theoretical study of Angus and Watts (1984) tried to predict the error present in the Bowen ratio method from the instrumental error. They found the combined relative error from net radiation and soil flux was around 4%. However this assumed (incorrectly) that ground heat flux is simply calculated from the soil heat flux. In reality ground heat flux involves a number of other measurements which increases error further. The paper deduces that for evaporation near the potential rate where $-0.2<\beta<0.2$, errors up to 30% in β produce errors of <5% in λE, therefore accuracy is good when evaporation is high. Seguin *et al.* (1982) suggested that the measurement errors of the Bowen ratio system were approximately 10% when all environmental and instrumental conditions were perfect and there were errors of 15-20% over periods greater than one month. Perrier *et al.* (1976) found that only 60% of Bowen ratio estimates of daily evapotranspiration lie within 20 % of actual measurements and 90% within 40%. Total evapotranspiration during the measurement period was found by summing the Bowen ratio estimates and this was found to be within 10% of actual evapotranspiration indicating that most of the errors were random. Recent improvements in instrumentation can reduce errors to 5%.

As this chapter contains no new experimental data and is theoretical in nature, its structure is slightly different from that of preceding chapters. The objectives are followed by an explanation of the approaches followed and then the main body of the chapter involves more detailed explanations of methods and example calculations. This main body is divided into two sections, the first focussing on the errors in the complete measured water balance and the second on the errors in the water requirements model. Finally the results are discussed and some conclusions drawn.

8.2 OBJECTIVES

The aim of error analysis is to estimate the accuracy of the water balance calculations. This is done by investigating how instrumental and measurement errors are propagated through the equations used in the water balance in order to quantify error rates of both the measured and modelled water balance. Errors are calculated first for the complete water balance, where all components are measured including evapotranspiration. Secondly an estimation of the errors in the water requirements model is calculated.

8.3 APPROACH

Two approaches were followed:

8.3.1 Mathematical

This addition method takes a pessimistic view and calculates the maximum possible error if all the errors are as large as they could be and there is no compensation. The following basic rules are applied following Topping (1962):

Addition and subtraction in equation – add ABSOLUTE uncertainties.

Multiplication and division in equation – add RELATIVE uncertainties.

Powers in equation – multiply relative uncertainties by the power.

The proof for this methodology is given using Taylor's expansion by Mannall and Kenwood (1994, p. 211-213).

8.3.2 Statistical

Statistical error analysis takes a more realistic view and is used to calculate the most likely errors. This requires replication, which was not always available, for example where there was only one set of meteorological instruments, or modelled data with no validation to compare it to. Replication was created using Monte Carlo analysis which uses random numbers to compute different scenarios of possible errors and calculate the most likely results. These can then be analysed statistically.

8.4 ERRORS WITHIN THE COMPLETE MEASURED WATER BALANCE

8.4.1 Bowen ratio

Instrument accuracy values are taken from the manufacturer's specifications:

- Vapour pressure measurement (General Eastern dew-10) ± 0.01°C
- Temperature measurement ± 0.01°C
- Soil heat flux plate ± 20%
- Soil moisture content (Delta T devices Theta probe) ± 1%
- Net radiometer (Kipp and Zonen NR lite) - There was no interpretable accuracy information available so the estimation of Angus and Watts (1984) of ± 2.5% was followed.
- Soil thermocouples ± 0.4% error in the difference between reference temperature and soil temperature (must be added to the reference temperature measurement)
- Reference temperature (within Campbell Scientific 21X datalogger) ± 0.1°C

Latent energy flux, and therefore evapotranspiration, is calculated from:

$$\lambda E = \frac{(R_n + G)}{(\beta + 1)} \tag{8.01}$$

where λE is latent energy flux, R_n is net radiation and G is ground heat flux and β is the Bowen ratio.

$\delta R_n = 0.025 \times R_n$ (0.025 is the relative error in R_n) (8.02)

where δ is the absolute error.

G is composed of two parts (a fact not considered in the analysis of Angus and Watts (1984)) and therefore is more complex than R_n.

$$G = G_{plate} + S \tag{8.03}$$

where G_{plate} is soil heat flux as measured by the soil heat flux plate and S is change in soil temperature above the soil heat flux plates

$$\delta G_{plate} = 0.20 \times G_{plate} \tag{8.04}$$

$$S = \frac{\Delta Ts \times (C_s z_s) + (418700 z_w)}{1200} \tag{8.05}$$

where ΔT_s is the change in soil temperature between the thermocouples (°C), C_{sw} is the specific heat capacity of water (4.187 MJ m^{-3} °C^{-1}), z_s is the depth of the soil layer being measured (0.08 m), z_{sw} is the depth of the standing water (m), Δt is the time interval (1200 s) and C_s is the specific heat capacity of soil (MJ m^{-3} °C^{-1})

$$\Delta Ts = Ts_n - Ts_{n-1} \tag{8.06}$$

$$\delta\Delta Ts = \delta Ts_n + \delta Ts_{n-1} \tag{8.07}$$

$$C_s = (1.93F_m + 2.51F_o + 4.187F_w)\times 10^6 \tag{8.08}$$

where F_m, F_o and F_w are the fractions of mineral, organic matter and water in the soil.

δF_w is from theta probe

F_m and F_o are estimated from soil component analysis (Appendix B). 95% confidence intervals can be used as repeated measurements were made.

$\delta F_m = \pm 5\%$ from 95% confidence interval

$\delta F_o = \pm 5\%$ from 95% confidence interval

$$\delta C_s = \delta F_m + \delta F_o + \delta F_w \tag{8.09}$$

$$\delta S = \frac{\delta\Delta Ts}{\Delta Ts} + \delta C_s + \delta z_w$$

The relative error of (Rn-G) is:

$$= \frac{\delta R_n - \delta G}{(R_n - G)} \tag{8.10}$$

The next step is to calculate the error in the Bowen ratio.

$$\beta = \gamma\frac{\Delta t}{\Delta e} \tag{8.11}$$

where Δe is the change in vapour pressure with height (kPa), ΔT is the change in temperature with height (K), and γ is the psychrometric constant (Pa K^{-1})

ΔT

$$\Delta T = T_l - T_u \tag{8.12}$$

$$\delta\Delta T = \delta T_l + \delta T_u \tag{8.13}$$

Δe

$$\Delta e = e_l - e_u \tag{8.14}$$

$$\delta\Delta e = \delta e_l + \delta e_u \qquad (8.15)$$

Bowen ratio

$$\delta\beta = \left(\frac{\delta\Delta T}{\Delta T} + \frac{\delta\Delta e}{\Delta e}\right) + \delta\gamma \qquad (8.16)$$

Latent energy

$$\delta\lambda E = \frac{\delta R_n - \delta G}{(R_n - G)} + \frac{\delta\beta}{\beta} \qquad (8.17)$$

The relative error in latent heat is equal to the relative error in evapotranspiration. Table 8.01 shows an example set of data.

Table 8.01 – An example of calculation of errors in the Bowen ratio approach for 01/06/01 at 12:40 (T_{ref} is reference temperature used in the calculation of T_s and T_{dew} is dew point temperature). The figures in bold typeface are those given by manufacturers, the others are calculated from these.

	Data	Error (%)	Error (absolute)
R_n (Wm^{-2})	279.87	**2.5**	6.99
G_{plate} (Wm^{-2})	15.206	**20**	3.04
$T_{s(n-1)}$ (°C)	11.972	**1.00**	0.12
T_s (°C)	12.253	0.97	0.119
ΔT_s (°C)	0.28113	84.822	0.24
T_{ref} (°C)	16.92	0.591	**0.1**
T - high (°C)	13.838	0.072	**0.01**
T - low (°C)	14.093	0.071	**0.01**
ΔT (°C)	0.25467	7.85	0.02
T_{dew} – low (°C)	15.81	0.32	0.05
e_d - low (kPa)	1.7952	0.56	**0.01**
T_{dew} - high (°C)	16.475	0.30	0.05
e – high (kPa)	1.873	0.53	**0.01**
T_{ref}-T_s (°C)	4.667	**0.40**	0.019

T_{ref}-$T_{s(n-1)}$ (°C)	4.948	**0.40**	0.020
γ(kPa °C^{-1})		**2.50**	
Δe (kPa)	0.0778	25.7	0.020
β	0.935	36.1	0.337
z_w (m)	1.88	1	0.019
F_o	0.20	5	0.01
F_m	0.91	5.00	0.05
C_w*z_w (MJ m^{-3} °C^{-1})	502	20.01	100541.69
C_s (MJ m^{-3} °C^{-1})	2.99	2.48	0.07
C_s+(C_w×z_w) (MJ m^{-3} °C^{-1})	502410	20.01	100541.76
S (Wm^{-2})	57.92	104.83	60.73
G (Wm^{-2})	73.13	87.19	63.77
R_n-G (Wm^{-2})	206.74	34.23	70.76
λE (Wm^{-2})	106.82	70.3	
ET (mm)	3.77	70.3	2.65

The maximum relative error in evapotranspiration is 70%. The reason for the large error is that big decreases in accuracy, in terms of relative errors, occur when subtracting nearly equal quantities. This occurs three times in the Bowen ratio method, when finding the difference in temperature and vapour pressure at two heights and the change in soil temperature over time. This is referred to as a loss of significance error. The greatest cause of error is in the change in soil temperature. The changes from one twenty-minute period to the next are often very small and due to the relatively low accuracy of the reference temperature, on which soil temperature's measurement depends, results in very large relative errors.

The sensitivity of the latent heat term to errors in ΔT_s was investigated (Figure 8.01). There is a linear relationship. An 84% error in ΔT_s as in Table 8.01 creates just a 2% error in latent energy.

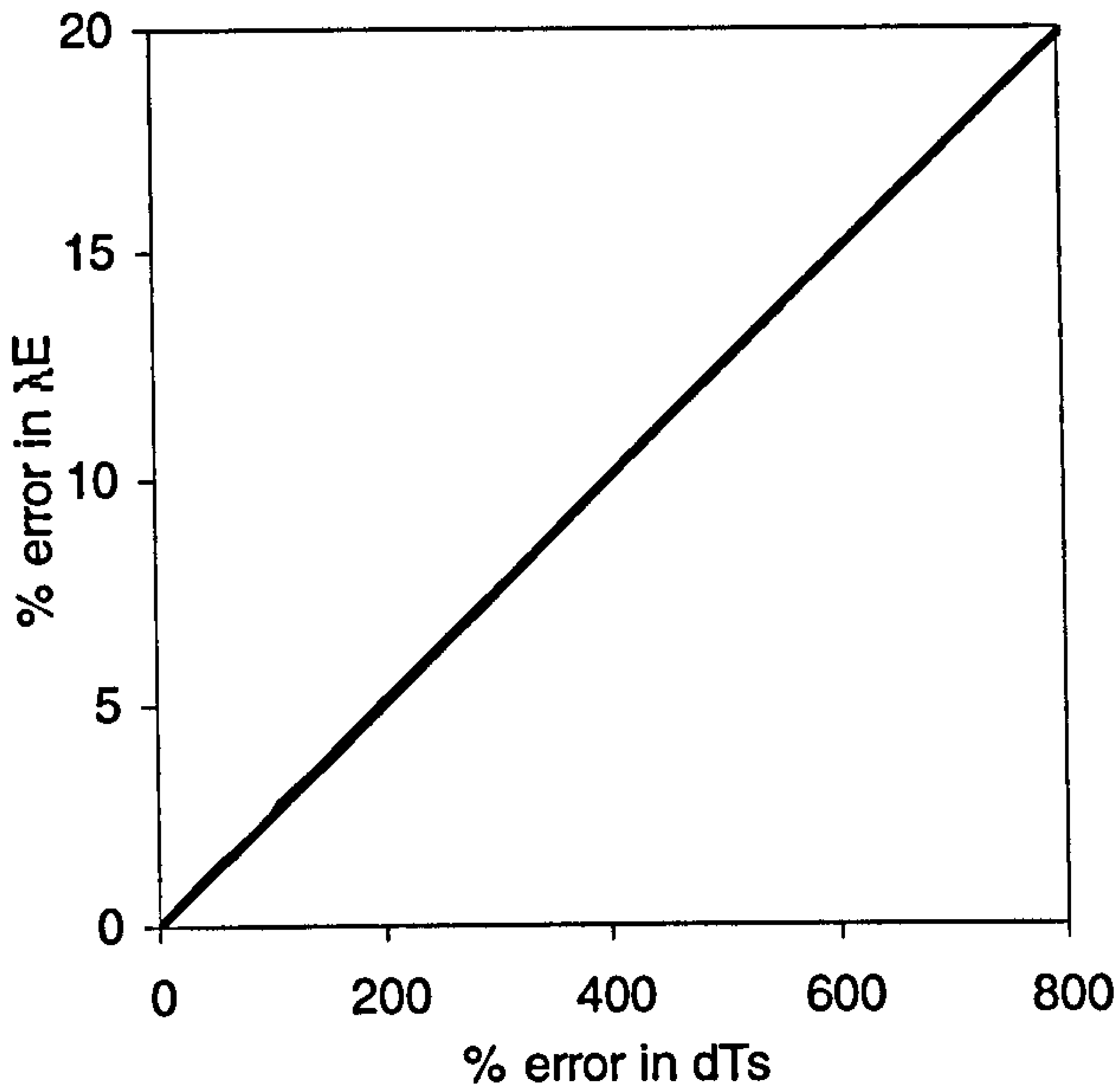


Figure 8.01 – Sensitivity analysis of the error in latent heat flux to the error in soil temperature change

Angus and Watts (1984) use an alternative method for working out the error in latent energy caused by the error in the Bowen ratio:

$$\lambda E + \delta\lambda E = \frac{(R_n - G)}{1 + (\beta + \delta\beta)} \tag{8.18}$$

$$\delta\lambda E = \frac{(R_n - G)}{1 + (\beta + \delta\beta)} - \frac{(R_n - G)}{1 + \beta} \tag{8.19}$$

$$\frac{\delta\lambda E}{\lambda E} = \frac{\dfrac{(R_n - G)}{1 + (\beta + \delta\beta)} - \dfrac{(R_n - G)}{1 + \beta}}{\dfrac{(R_n - G)}{1 + \beta}} \tag{8.20}$$

$$\frac{\delta\lambda E}{\lambda E} = \frac{1 + \beta}{1 + \beta + \delta\beta} - 1 \tag{8.21}$$

$$\frac{\delta\lambda E}{\lambda E} = \frac{\delta\beta}{1 + \beta + \delta\beta} \tag{8.22}$$

Using the data in Table 8.01:

$$\frac{\delta\lambda E}{\lambda E} = \frac{\delta\beta}{1 + \beta + \delta\beta} = \frac{0.337}{1 + 0.935 + 0.337} = 0.148 \tag{8.23}$$

This results in an overall error of 14.8%, which is much smaller. This different equation is derived due to the omission of second order terms in the derivation, although no justification for this is given.

The above methods predict the worst case scenario when all errors are at their maximum and none cancel one another out. Monte Carlo analysis was used to find the most likely errors using the program Crystal Ball (Decisioneering Inc., Colorado, version 3.0). This was done by generating random numbers within the specified limits (both positive and negative) to describe the level of error in each measurement. It was assumed that all errors had a uniform distribution and a large error was as likely as a small one. This process was repeated 50 000 times for the above example data set and the results can be seen in the frequency histogram (Figure 8.02).

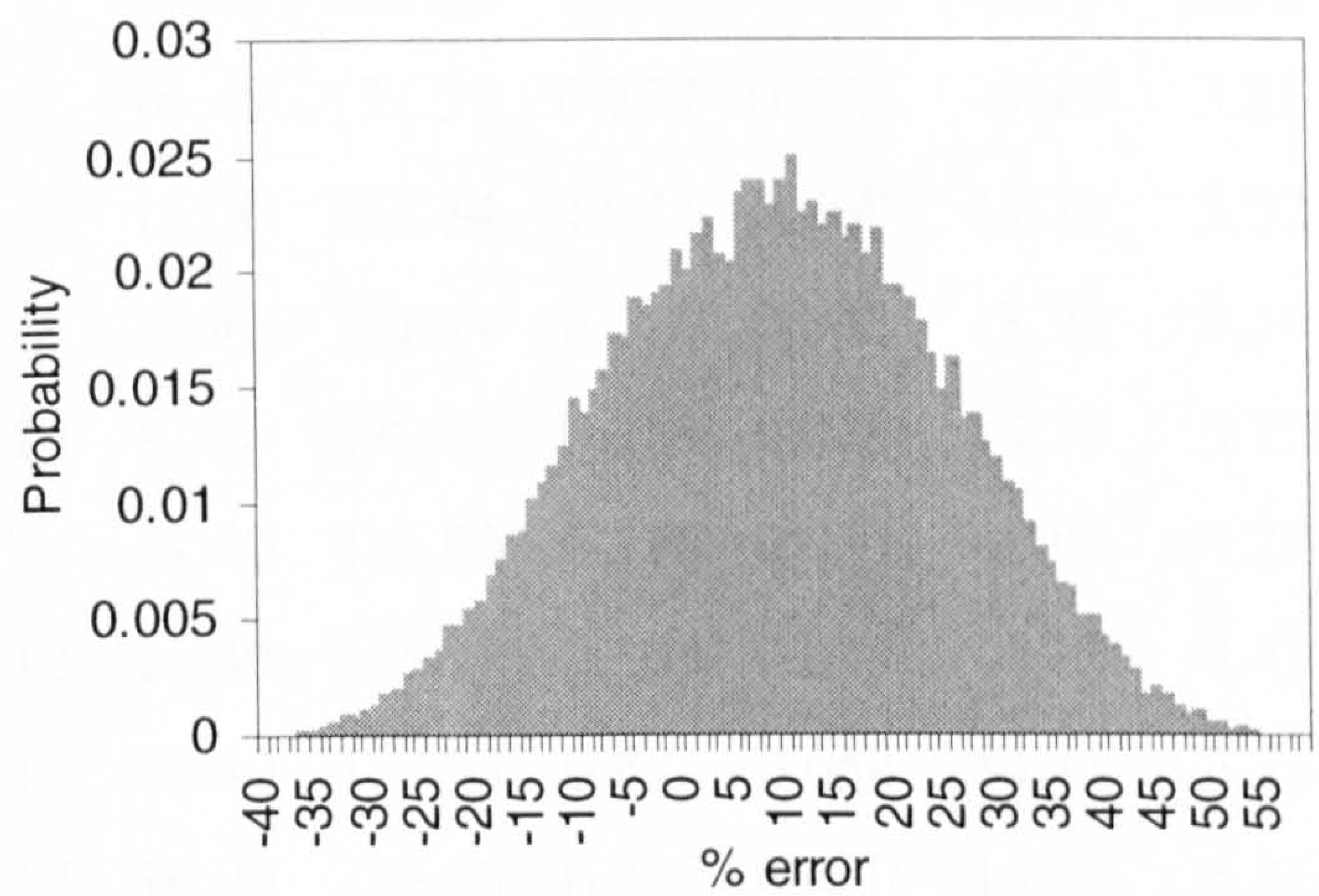


Figure 8.02 – Probability histogram of errors within the Bowen ratio approach using data from 01/06/01 at 12:40

The histogram shows that the most probable error for this data set is 9.5% with a standard deviation of 15.9. The maximum error was around 60% - the theoretical error of 70% above was never achieved even in 50 000 runs.

The errors are different with different data sets and therefore Monte Carlo simulations were carried out on 23 example data sets. The results are in Table 8.02 in order of increasing evapotranspiration. The mean relative error is given with the standard deviation of the frequency histogram. The absolute error at one standard deviation is given – there is a 68% chance that the absolute error will be less than this value.

Table 8.02 – Data and error results for 23 example sets of Bowen ratio measurements

Run	Time	R_n (Wm^{-2})	ΔT (°C)	Δe (kPa)	G (Wm^{-2})	ET (mm)	Mean rel. error (%)	SD	Abs. error at 1 S.D. (mm)
1	19:20	7.98	0.611	0.163	-17.89	0.73	2.7	98.8	0.72
2	5:40	2.91	0.375	0.098	-14.38	0.82	-21	412.8	3.38
3	19:00	61.84	0.842	0.039	-14.55	1.10	5.2	39.5	0.43
4	13:20	165.79	0.896	0.058	19.35	2.54	8.7	22.4	0.57
5	6:40	55.71	0.276	0.028	-68.05	2.62	8.6	56.1	1.47
6	13:00	685.82	3.437	0.036	69.96	2.92	8	25.2	0.74
7	16:40	131.56	0.237	0.033	-3.95	3.21	3.4	31.5	1.01
8	15:00	328.84	1.928	0.068	39.79	3.52	6.4	14.8	0.52
9	16:40	177.14	0.315	0.038	18.94	3.59	8	27.1	0.97
10	12:40	279.87	0.255	0.078	73.13	3.77	9.5	15.9	0.60
11-	17:00	215.83	0.376	0.033	0.95	4.29	2.6	27.6	1.18
12	15:40	143.70	0.105	0.039	-4.75	4.43	3.1	28.5	1.26
13	12:40	342.98	0.673	0.054	54.11	5.56	6.6	17.5	0.97
14	16:00	223.28	0.237	0.070	19.58	5.86	7.3	17.4	1.02
15	17:00	198.67	0.017	0.040	18.65	6.17	6.5	54.3	3.35
16	11:00	547.74	1.493	0.083	83.94	7.42	7.9	11.3	0.84
17	13:40	494.36	1.062	0.072	53.56	7.81	5.8	12.8	1.00
18	12:00	565.91	1.084	0.083	57.11	9.57	6.2	11.1	1.06
19	14:00	452.52	0.463	0.065	47.40	9.66	5.8	14.2	1.37
20	12:40	619.03	0.436	0.080	62.52	14.37	5.3	11.3	1.62
21	13:20	612.90	0.387	0.081	58.32	14.82	4.8	11.3	1.68
22	17:20	181.17	0.489	0.054	6.06	15.55	6.3	21	3.27
23	13:00	634.79	0.349	0.089	61.90	15.99	4.9	10.5	1.68

The probability histograms for the likely errors within some of these data sets can be seen in Figure 8.03.

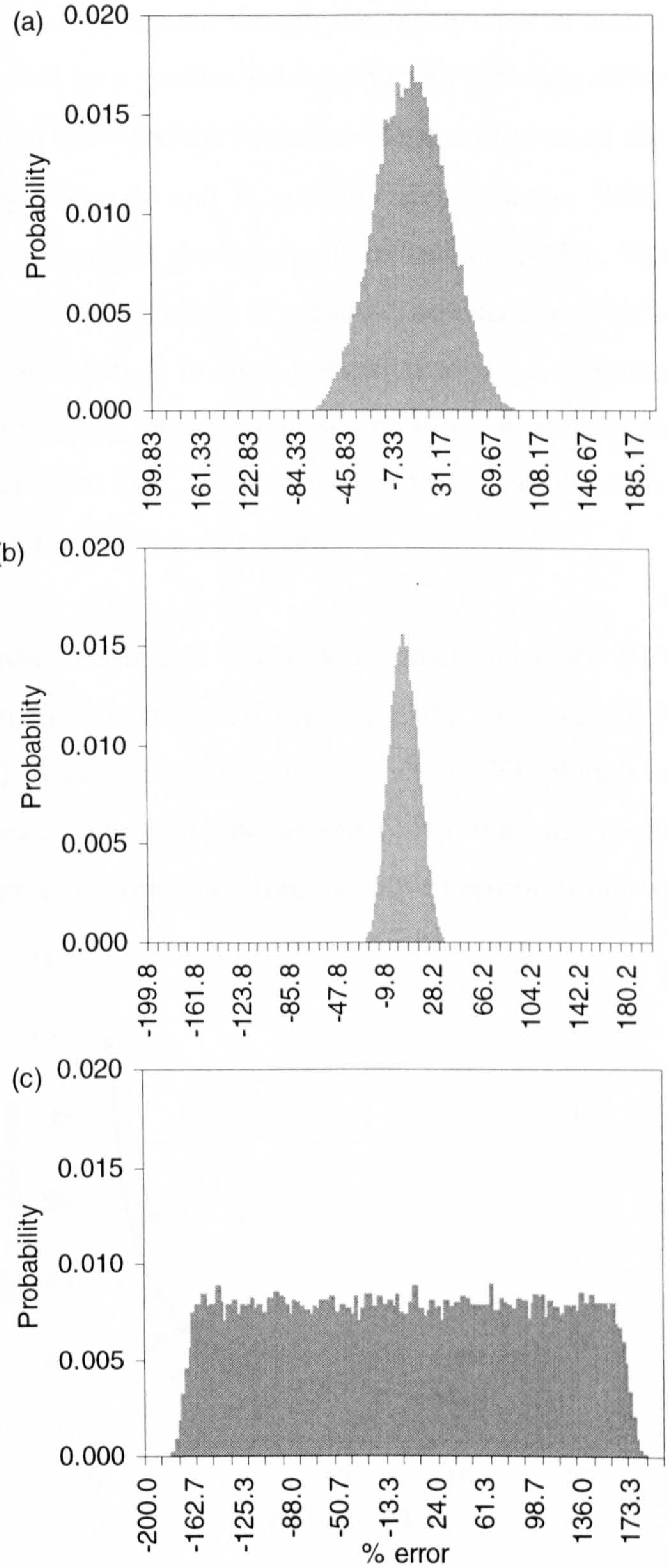


Figure 8.03 — Probability histogram showing the probability of errors of different magnitudes occurring with different data sets (a) run 9, (b) run 23, (c) run 1.

In run 9 the standard deviation is larger than in the initial example giving a larger range of possible errors, though the mean error is similar. This larger standard deviation is caused by a smaller value of vapour pressure difference. Run 23 is typical in terms of mean and standard deviation of error of most of the examples tried in the middle of the day when Δe and R_n are sufficiently large. With this type of data we can be 95% confidant that the error will be less than 25%. The majority of data sets resulted in a normal distribution of means. The exceptions as in the case of run 1 (Figure 8.03c), occur when G is large compared to R_n. As demonstrated when calculating maximum errors, G is an important source of error, usually controlled due to its small magnitude compared to R_n. However when this is not the case as in the example above, errors can quickly become very large.

Errors are largest when Δe is small and when R_n is small compared to G. Both these situations occur most often in early morning and late evening when the magnitude of ET is also small. Figure 8.04 shows ET plotted against the standard deviation of the mean error from the Monte Carlo analysis. It can be seen that the largest standard deviations (and therefore the largest errors) occur when evapotranspiration is small.

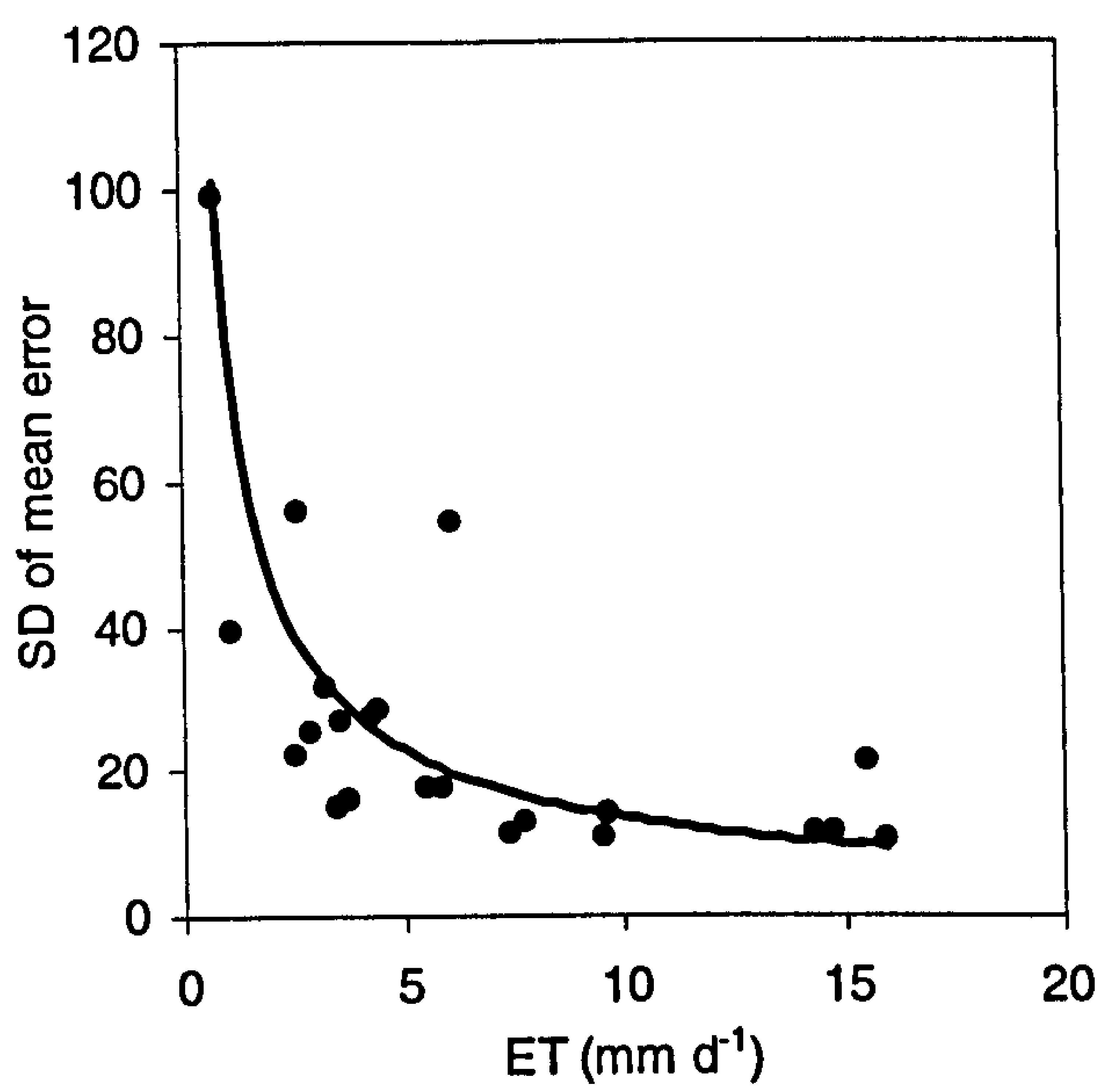


Figure 8.04 - Evapotranspiration plotted against the standard deviation of the probability distribution of the error in the Bowen ratio approach as derived from Monte Carlo analysis.

The fact that the largest relative errors occur with the smallest measurements of evapotranspiration means that the absolute error will also remain small at these times. This is demonstrated in Figure 8.05. It can be seen that, in general, absolute error increases with evapotranspiration measurement. There are three points that do not fit into this pattern.

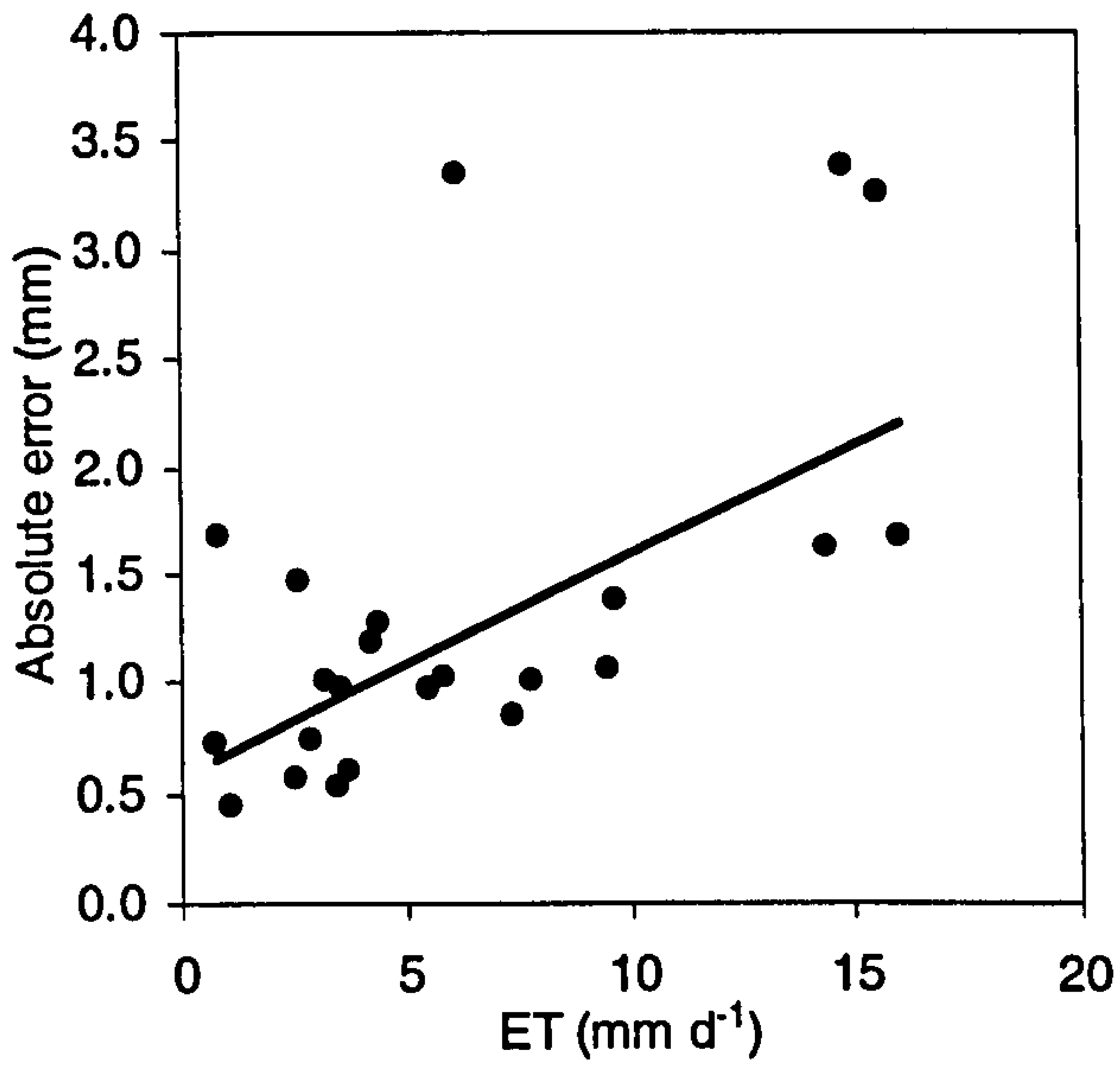


Figure 8.05 – The relationship between absolute error in evapotranspiration and evapotranspiration magnitude.

The above analysis is for mean twenty-minute measurements. However the water balance uses daily averages which will result in a possible reduction of the magnitude of errors due to the cancelling out of under-estimation and over-estimations. The effect of this was investigated by running the Monte Carlo analysis for just 42 simulations (representing 14 hours of daylight). This was repeated 20 times and it was found that the average and standard deviation of these runs was very similar to that of 50 000 runs as shown in the two examples in Table 8.03.

Table 8.03 – Comparison of the results of 50 000 runs of the Monte Carlo simulation with 42 runs representing one day.

	50 000 runs	42 runs × 20
Mean	9.5	8.755
S.D.	15.9	15.225
Mean	6.5	6.445
S.D.	54.3	54.475

The other complication is that there is a whole range of different data sets which are measured each day, each with their own error range so it is difficult to pick out a single value to represent the error for a daily estimate.

8.4.2 Streamflow measurement

The manufacturers of the Doppler flow meter give an accuracy for the Starflow instrument of ± 2% for velocity measurement and 0.25% for depth measurement giving a total error for discharge of 2.25%.

8.4.3 Rainfall measurement

There are a number of potential sources of error within rainfall measurements (Winter and llamas 1993). The instrument error for the type of automatic tipping bucket raingauge used by the Environment Agency is ± 1%. In addition error can result from improper use of windshields and placements of gauges. These can be in the range of 5-15%. An even greater problem is in determining total areal distribution from point data. In the area of study there is roughly a density of 18 km^2 per gauge. This results in around a 10% error (Winter and llamas 1993). An estimation of the total error in rainfall measurements may be around 16%, assuming the placement of gauges by Environment Agency professionals is reasonable.

8.4.4 Storage

Sources of error in calculating storage volume on Stodmarsh NNR are:

- In the creation of the average elevation above datum. Depth measurements do not always occur exactly in the same places as levelling measurement and because the surface is not flat there is potential for error over quite short distances.

- Due to the inaccuracy of the levelling survey which is only measured to the nearest 0.1 m.
- In conversion of depth to volume due to possible inaccuracies in the measurement of the site area.
- Within the error in the regression line when gaugeboards are used (Figure 5.11).

The equation used to create storage volume (V) from mean elevation above datum ($\bar{y}$) is:

$$V = \left(\frac{\sum (\bar{y} - L)}{n} \right) \times A \qquad (8.24)$$

where A is the site area (m^2), L is the level above Ordnance datum as recorded in a levelling survey prior to this research and n is the number of levelling points used. There are errors in $\bar{y}$, L and A all of which must be included in the final error total.

To work out the error in $\bar{y}$, 95% confidence limits were created on the average elevation above datum measurements. It is possible to say that we are 95% confident that the depth falls within this range. Absolute and relative errors were created as the difference between the 95% limits and the mean.

The original levelling survey was only carried out to the nearest 0.1 m. The impact of this inaccuracy was investigated using Monte Carlo analysis. For each levelling point a random number within ± 0.05m of the level is used. This was repeated 50 000 times and again 95% confidence intervals of the resulting average level were found.

The area measurement is to the nearest hectare. The new area is 79 ha ± 0.5. This results in a possible error of 0.63%.

The total relative error in storage measurements is:

$$\frac{\delta V}{V} = \frac{\delta \bar{y} + \delta L}{(\bar{y} - L)} \times \frac{\partial A}{A} \qquad (8.25)$$

These calculations can be seen in Table 8.04.

Table 8.04 – Calculations of total error in measured storage elevations

Elevation (m)			0.712	0.747	0.782	0.829	0.854	0.888	0.766
Measurement errors	SE		0.011	0.005	0.011	0.011	0.011	0.014	0.01
	95% con.	+/-	0.029	0.016	0.048	0.03	0.029	0.036	0.031
	% error		4.094	2.077	6.136	3.674	3.378	4.055	4.036
Levelling errors (δL)	Mean level	(m)	0.543	0.555	0.568	0.586	0.595	0.601	0.562
	95% con.	+	0.006	0.004	0.007	0.005	0.004	0.006	0.006
		-	0.007	0.005	0.007	0.007	0.004	0.006	0.006
	% error	+	1.052	0.782	1.199	0.927	0.733	0.982	1.156
		-	1.269	0.837	1.185	1.137	0.683	0.932	0.984
Area errors (δL)		%	0.63	0.63	0.63	0.63	0.63	0.63	0.63
Total (δV)		%	21.2	11	26.2	15.8	13.4	15.2	18.9

For a mean elevation above datum of 0.829 m as measured on 07/12/00 the errors are calculated using the above data as:

$$\frac{\delta V}{V}=\frac{\delta\bar{y}+\delta level}{(\bar{y}-level)}\times\frac{\partial A}{A}=\frac{0.030+0.007}{0.829-0.586}\times\frac{0.5}{79}=0.158=15.8\% \qquad (8.26)$$

When gaugeboard readings are used, the error analysis is as above, with the exception of the method calculation of the error in $\bar{y}$. In this case, $\bar{y}$ is found from the error of the regression line which is calculated from the standard errors of the estimates, allowing confidence limits around the regression line to be established (Figure 8.06).

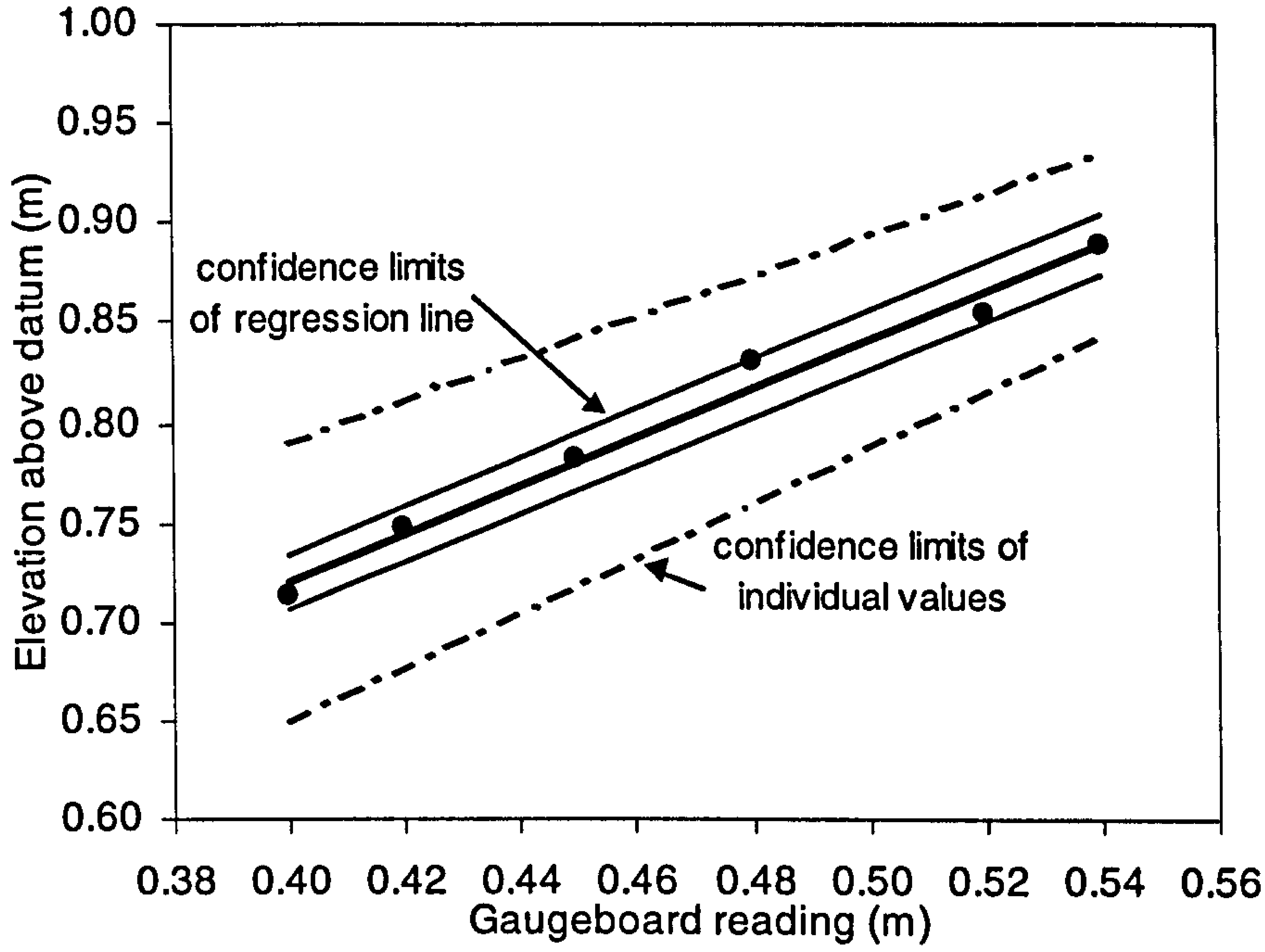


Figure 8.06 – The relationship between gaugeboard readings and measured elevation above datum, with best fit line (solid bold line), the confidence limits of the regression line (solid lines) and the confidence limits of individual values (dashed lines).

The inner confidence zones represent the confidence limits for the regression line. The outer confidence limits are those for individual predicted values of *y* from *x*. These predicted values are subject to an additional source of error, which is scatter about the regression line (Fowler *et al.* 1998). This represents a considerable extension to the width of the confidence zones. These outer confidence limits are used to find the 95% confidence interval for the calculations from the gaugeboard. This resulted in smaller relative errors for the larger gaugeboard readings according to the reduced confidence limits that can be seen in the graph above. The errors ranged from 7.4% for a gaugeboard reading of 0.39 m to 3.6% for a reading of 0.54 m.

The complete measured water balance quantified storage using two gaugeboard readings for the start and the end of the period. The errors for these readings are calculated in Table 8.05.

Table 8.05 – Calculation of the errors in the storage depth estimation for the start and end of the period over which the complete measured water balance was measured.

		30/05/01	22/08/01
Gaugeboard	x	0.42	0.45
Elevation	y	0.744	0.780
Volume (m^2)		201339.6	239829.5
Regression errors			
	xy	0.313	0.351
	y^2	0.554	0.609
	x^2	0.176	0.203
Sums of Products		0.018	0.018
Sums of Squares y		0.0223	0.0223
Sums of Squares x		0.0153	0.0153
Residual variance		8.78×10^{-5}	8.78×10^{-5}
	SE	0.017	0.016
95% confidence	+	0.792	0.824
	-	0.696	0.737
Absolute error		0.048	0.044
% error		**6.453**	**5.600**
Levelling errors			
Mean level		0.554	0.567
95% confidence	+	0.559	0.574
	-	0.549	0.560
Absolute error	+	0.005	0.007
	-	0.005	0.007
% error	+	**0.821**	**1.167**
	-	**0.890**	**1.176**
Area			
% error		0.630	0.630
Total %		**28.283**	**24.243**
δV		**56943.97**	**58140.82**

The total errors in each of the gaugeboard measurements are 28% and 24%. These are measurements of storage at a single point in time. The input to the water balance is the change in storage over time, in this case the difference between the two measurements above. The absolute errors in the calculated storage volumes to which the above gaugeboard readings correspond must be summed, in order to find the total relative error in change in storage.

$$\%error = \frac{\delta V + \delta V}{V - V} = \frac{56943.97 + 58140.82}{239829.5 - 201339.6} = 2.99 \qquad (8.27)$$

Because of the relatively small change in storage over this period this results in a 299% error which is very large compared to the individual errors of each measurement and particularly the errors caused by the regression line, levelling measurements and area measurements.

8.4.5 The measured water balance

To estimate the error in the complete water balance the absolute errors of each component are simply added together. The errors are summarised in Table 8.06.

Table 8.06 – A summary of the errors present in the complete measured water balance

	% Error	Volume	Absolute Error
Rain	16	407829.1	65252.66
Stream in	2.25	397825.7	7956.514
ET	8	439348.6	35147.89
Stream out	2	337632.1	6752.642
Storage	299	38489.89	115084.8
Sum		1621125	230194.5
Overall % error	**14**		

The total error in the measured water balance is 14%.

8.5 ERRORS WITHIN THE MODELLED WATER BALANCE

Errors were also quantified within the water requirements model. This model is based primarily on the balance between daily rainfall and evapotranspiration and also includes stream inflow as an indication of water resource availability. Errors were calculated for the summer (April to September) water balance in order to determine the levels of confidence that can be placed on estimates of the quantity of water required from the Lampen Stream to compensate for on-site deficits. Due to the uncertainties in quantifying errors in a model, a Monte Carlo approach was used (again using the software Crystal Ball), to determine the most likely range of errors. Estimated error distributions were defined for each daily evapotranspiration and rainfall datum. The model was run 2500 times for each summer data set, using random combinations of error and a frequency chart of the 2500 resulting estimates of water requirement was plotted, allowing means and standard deviations to be calculated.

8.5.1 Rainfall

Rainfall is not modelled, as historical data is available and therefore errors are assumed to remain consistent with those in the measured water balance. Errors were estimated as up to 16% on a daily basis. It was assumed that the errors have a uniform distribution (as there is no way of determining the actual distribution) implying that any error is as likely as any other within the range –16% to +16%.

8.5.2 Evapotranspiration

The errors in the evapotranspiration model were quantified by comparing the daily results of the model created using Penman Monteith reference evapotranspiration using data from Manston airport and crop coefficients from the Walton Lake site, Buckinghamshire as described by Fermor *et al.* (2001), with the measured Bowen ratio evapotranspiration data collected in 2001 and 2002. The 2001 Bowen ratio data were used to calibrate the model, but the two data sets are entirely independent and its inclusion increases the number of days available for comparison (139 days were used over the two seasons). The percentage error was calculated on a daily basis and plotted as a probability histogram (Figure 8.07).

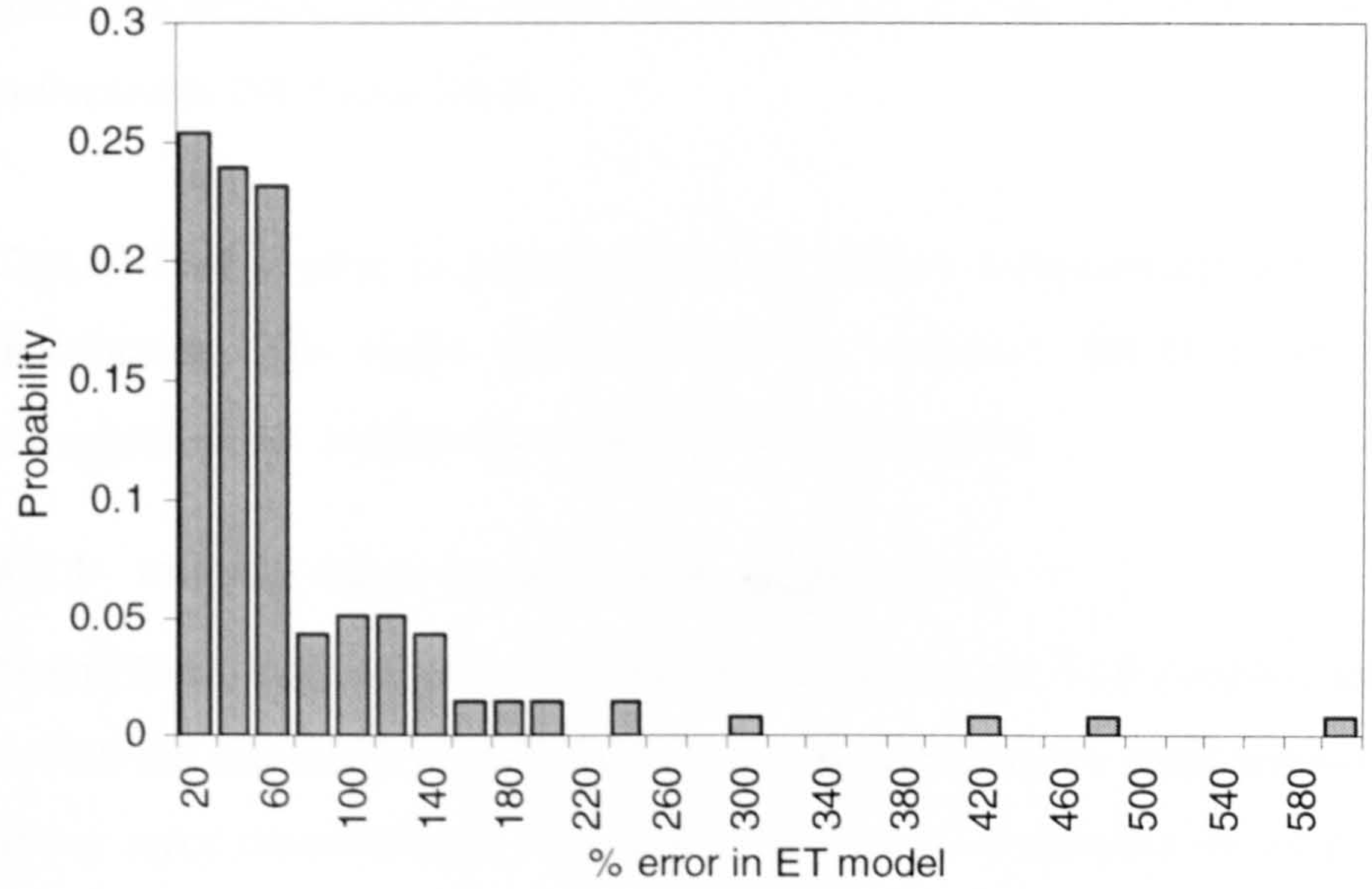


Figure 8.07 – Probability distribution of the percentage error in daily ET estimations created using Penman Monteith reference evapotranspiration using data from Manston airport and crop coefficients from the Walton Lake site, Buckinghamshire as compared with the measured Bowen ratio evapotranspiration data collected in 2001 and 2002.

The lognormal distribution is widely used in order to normalise data where values are positively skewed, as in Figure 8.07. The normalised data is re-plotted in Figure 8.08.

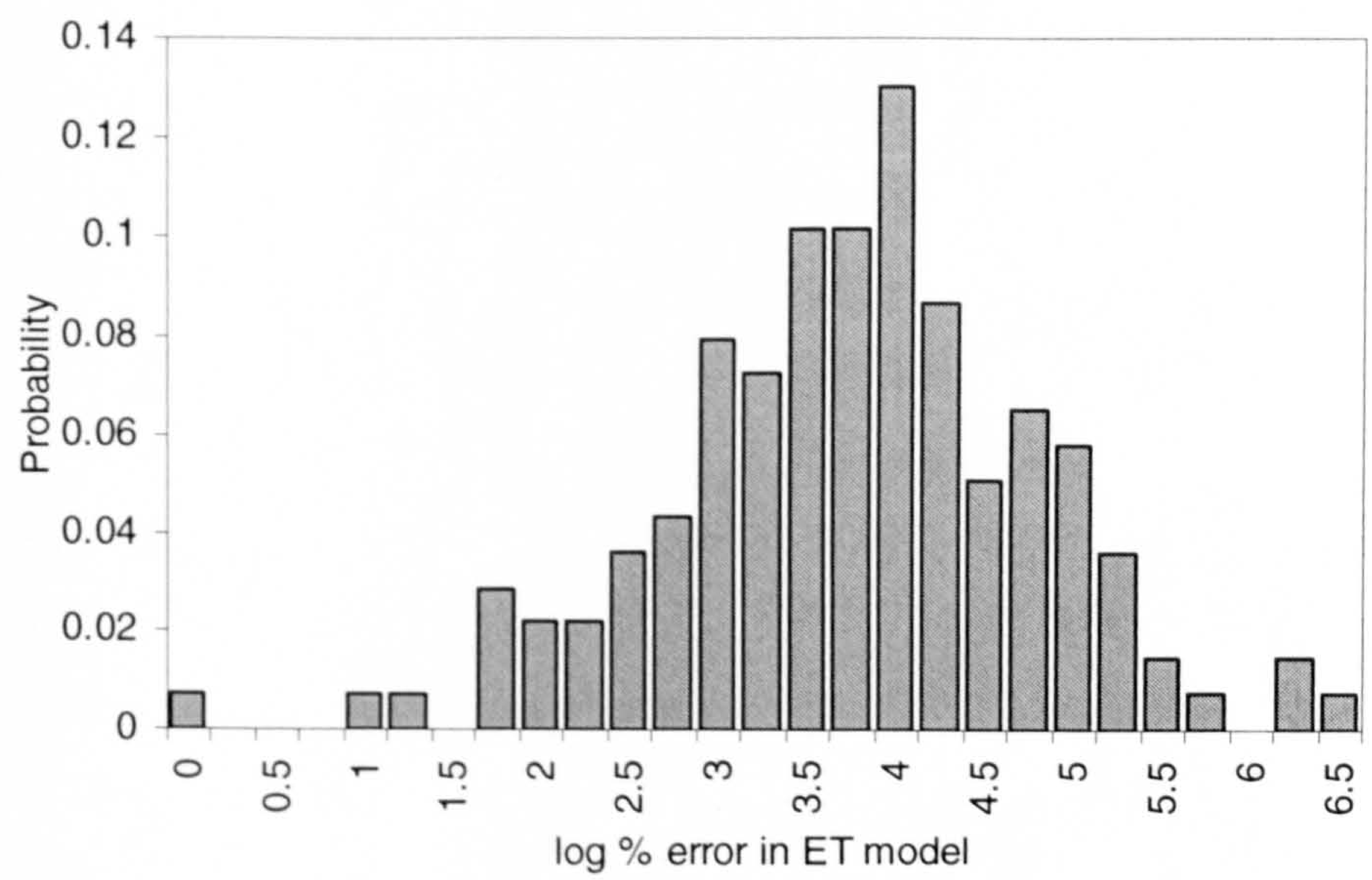


Figure 8.08 – Probability distribution of log of percentage error in daily ET estimations, created using Penman Monteith reference evapotranspiration using data from Manston airport and crop coefficients from the Walton Lake site,

Buckinghamshire, compared to the measured Bowen ratio evapotranspiration data collected in 2001 and 2002.

Crystal Ball is able to generate random values representing errors over a log normal distribution. The mean (63.6%) and the standard deviation (81.9) of the original unlogged errors defined the range of the distribution.

8.5.3 Overall water requirements model error

Monte Carlo simulations were carried out within the water requirements model in order to find out the range of possible water requirements for each summer of data, using the above error distributions. Figure 8.09 gives two examples of frequency histograms of possible water requirements over individual summers. The two years chosen are representative of the main types of the resultant distributions. Those years that require a large amount of water form a normal distribution. Those with requirements for little or no water have a very strongly positively skewed distribution, as it is impossible for the value to fall below zero.

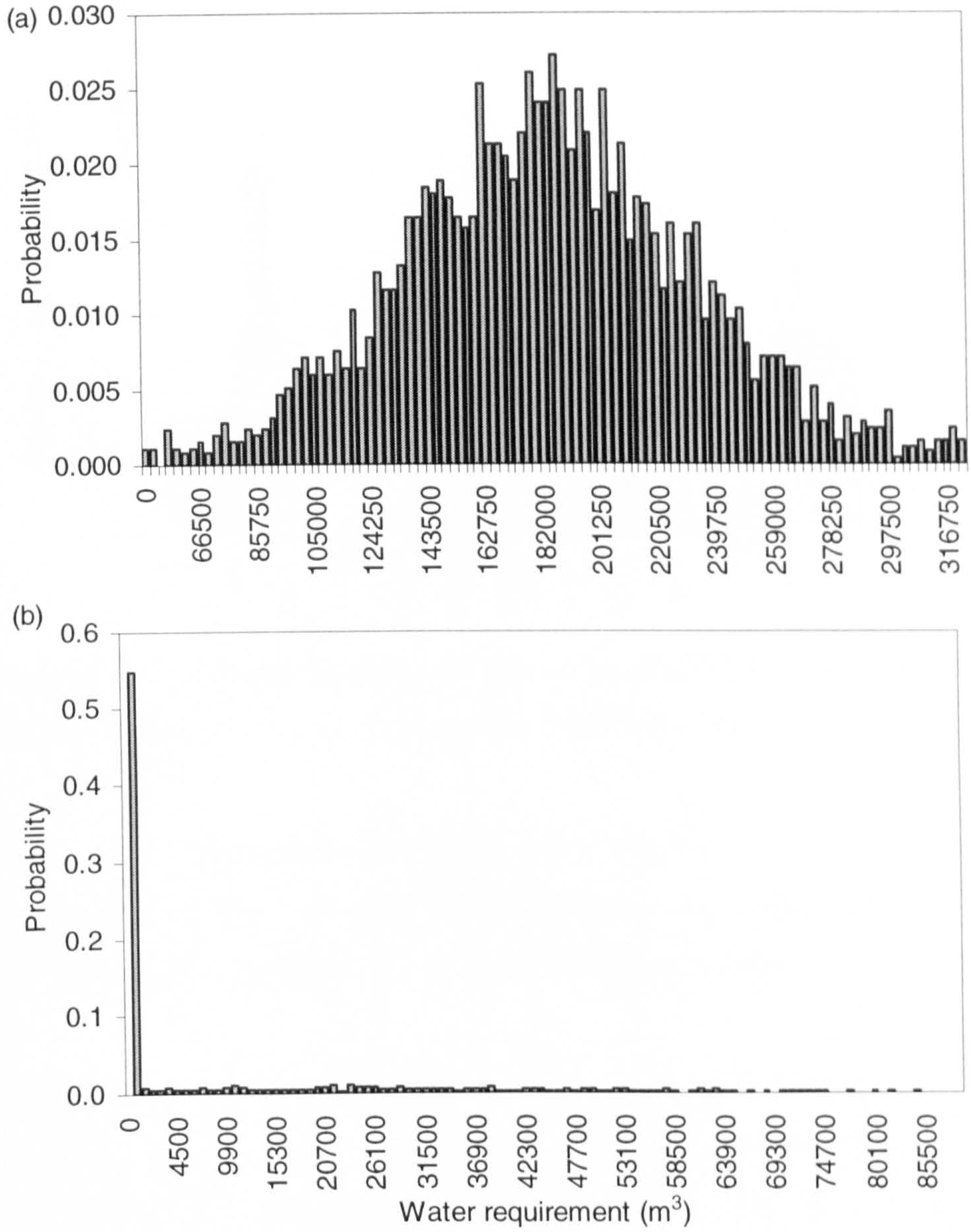


Figure 8.09 – Distribution of water requirements with Monte Carlo error analysis for summer (a) 1975 and (b) 1999.

The standard deviations were calculated from the probability distributions of the water requirements and from these the range within which we are 95% confident the actual value for that set of data will fall (two standard deviations of the mean) was found. These ranges were plotted as error bars in Figure 8.10.

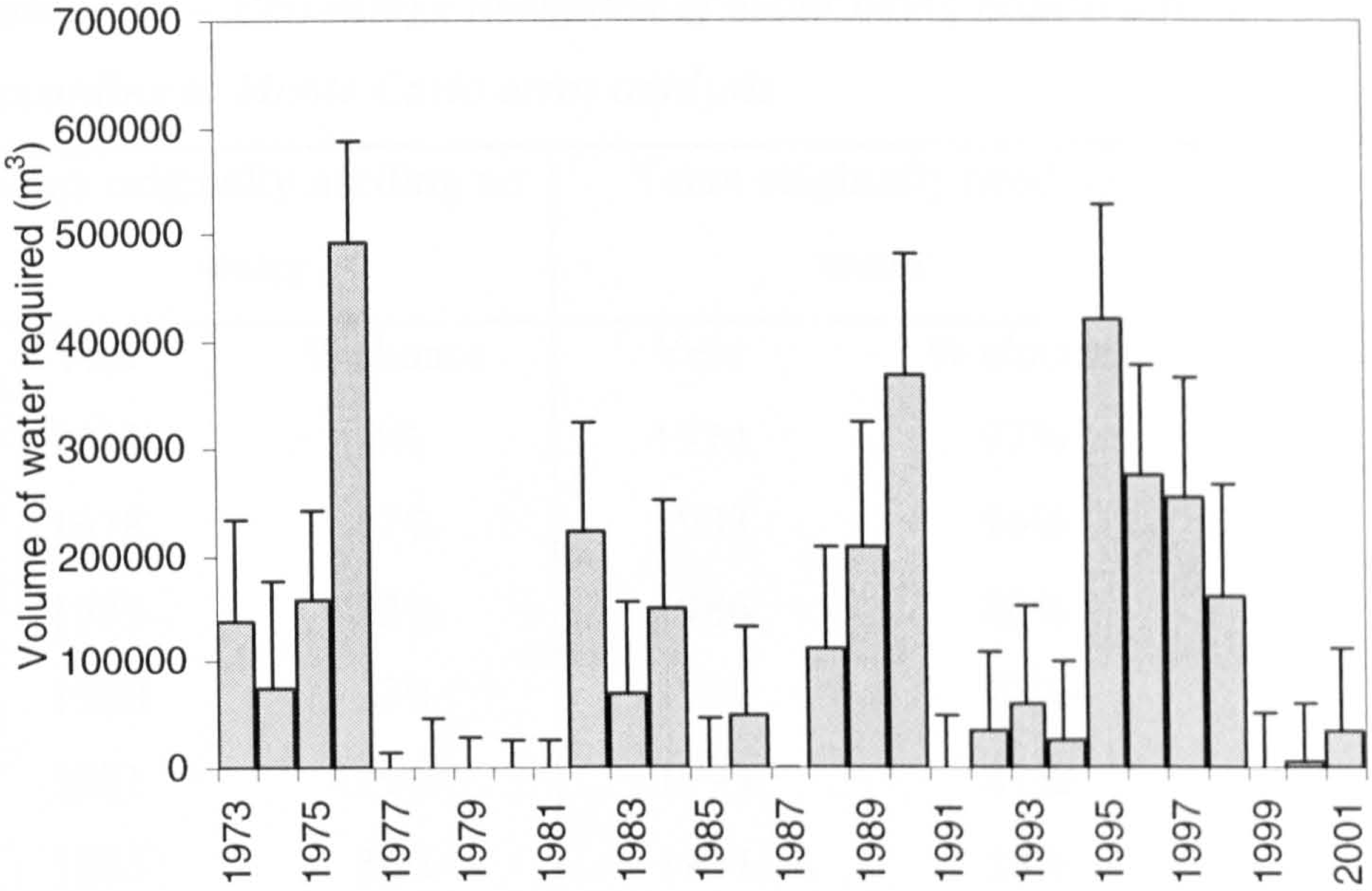


Figure 8.10 – Modelled volume of water required from the Lampen Stream for each summer with error bars showing 95% confidence ranges for the data.

The size of the standard error bars is fairly similar from year to year. This means that as a proportion of the estimated water requirement, the uncertainty is smaller in years with a large water requirement and larger in years with a small water requirement.

There is uncertainty caused by the fact that although the original model determined that 9 out of 29 years would require no water, some combinations of error reveal that water may be required in these years, as demonstrated by Figure 8.09b. Of the nine years that were originally deemed to require no water from the Lampen Stream, only one year would not require any water under any combinations of error. This was 1987, the year with the highest rainfall total. In a similar way some years deemed to have low water requirements could in fact have none. The percentage likelihood of these years requiring water from the Lampen Stream is shown in Table 8.07.

Table 8.07 – Percentage likelihood of water being required from the Lampen Stream according to Monte Carlo error analysis

Years originally needing no water		Years originally needing water	
Year	% chance	Year	% chance
1977	9%	1974	97%
1978	47%	1983	96%
1979	22%	1986	86%
1980	23%	1992	77%
1981	19%	1993	87%
1985	36%	1994	33%
1987	0%	2000	56%
1991	75%	2001	79%
1999	55%		

The original model estimated that water would be required from the Lampen Stream in 69% of years. If the 17 data sets in Table 8.07 are each assumed to represent 100 years, with water required in the number of years as determined by the percentages, then this results in water being required in 897 out of 1700 years. If it is assumed that the other 12 years not listed above have water requirements every year (1200 out of 1200 years, again assuming each annual data set represents 100 years) then this results in a total requirement for water in 2097 out of 2900 years, which is 72% – close to the original value of 69%.

From the 95% error bars in Figure 8.10 it is possible to redraw the probability chart showing the percentage of water required from the Lampen Stream used in Chapter 6 with 95% confidence intervals.

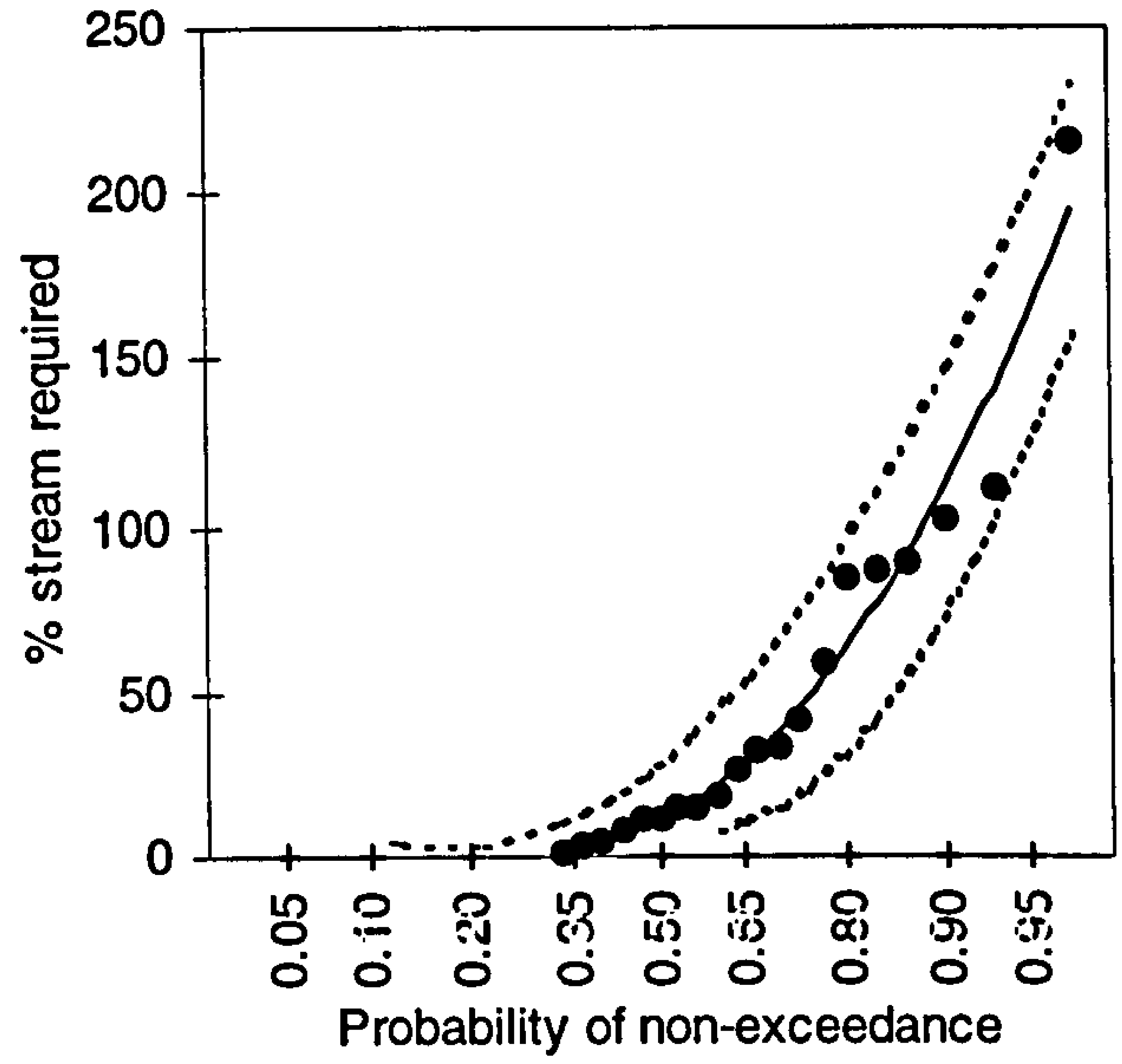


Figure 8.11 – The probability of requiring percentages of water from the Lampen Stream with 95% confidence limits calculated from two standard deviations of the range of the Monte Carlo-generated water requirement frequency histograms.

The range of error gets narrower as the demand for water becomes lower. There is a potential range of ±~0.14 in the probability of exceeding a demand for 25% of the Lampen Stream, a range of ±~0.11 when determining whether more than 50% is required and a range of ±~0.08 in the probability of requirements exceeding 100% (Table 8.08).

Table 8.08 – The maximum, minimum and mean probability of requiring more than 25%, 50% and 100% of the Lampen Stream

	Min.	Mean	Max.
>25% inflow	0.22	0.36	0.50
>50% inflow	0.17	0.27	0.39
>100% inflow	0.08	0.14	0.22

The above analysis of the errors in the water requirements model assumes that the modelled inflow data is correct. However the IHACRES inflow model must also be analysed for errors. The model output is used solely on a seasonal basis and therefore errors can be calculated over this period rather than on a daily basis as above. This should result in a reduction of the error level, as errors should compensate within the model. Unfortunately the inflow model only has a very short validation period – there are only 3 months of complete data. The monthly and seasonal errors are in Table 8.09

Table 8.09 – Monthly and seasonal error in the calibration of the IHACRES inflow model

Month (2002)	Measured total streamflow ($m^3 d^{-1}$)	Modelled total streamflow ($m^3 d^{-1}$)	Moduli of % error
July	49210.8	47494.31	3.5
August	98681.3	95925.35	2.8
September	102937.2	92052.2	10.6
Seasonal	147892.1	143419.7	7.8

In order to estimate the maximum effect of this error, the percentage of the Lampen Stream was calculated from the maximum water requirements (+ 2 S.D.) and the minimum Lampen Stream volume (-7.8%), and the minimum water requirements (- 2 S.D.) and the maximum Lampen Stream volume (+7.8%), to simulate the most extreme conditions. This was then plotted on the probability graph as a further set of confidence limits (Figure 8.12).

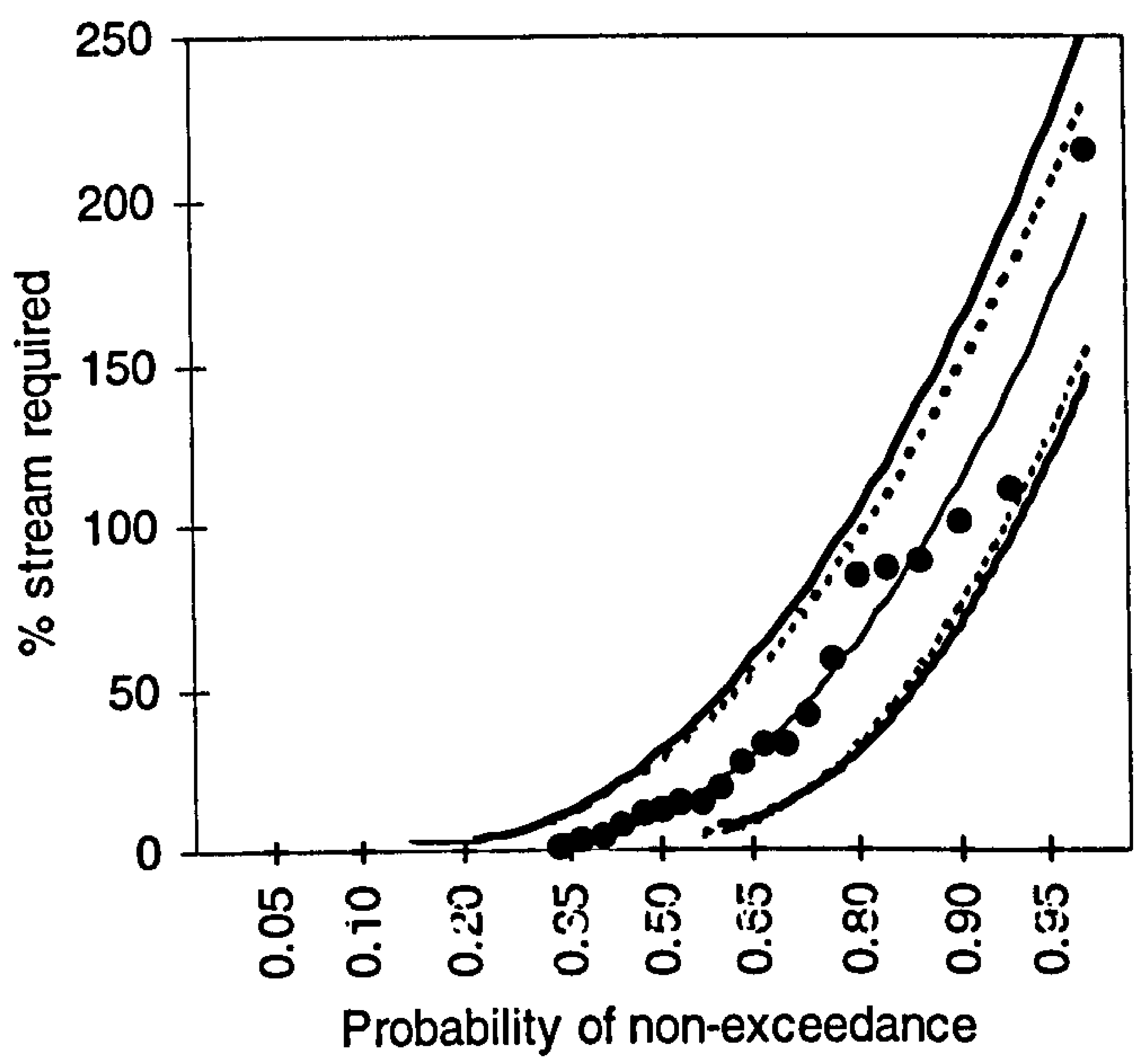


Figure 8.12 – Modelled probability of water requirements from the Lampen Stream (data points with trendline) and confidence intervals due to error in water requirement

modelling (dashed lines) and confidence intervals due to a combination of error in requirement modelling and streamflow modelling (solid lines).

The overall error due to error in the IHACRES streamflow model is small compared to that caused by uncertainty in water balance modelling due to that fact that on a seasonal basis errors compensate and become smaller than those on a daily basis.

8.6 DISCUSSION

As accurately quantifying a water balance is difficult, so too is accurately quantifying the errors in that water balance, as to a large extent they must be based on assumptions and probabilities. Decisions must also be made as to whether it is most useful to know the worst case scenario in terms of error, or the most likely event.

A number of previous studies have been carried out in order to assess the accuracy of the Bowen ratio technique, though none have been as comprehensive as the above calculations in assessing all the possible instrumental error involved in the technique. As it is a complex technique with many inputs and parameters, there are many potential sources of error. However the largest errors are caused by the fact that the technique relies on finding the difference in soil temperature, air temperature and dewpoint temperature over quite small distances. This results in small values for the difference, whose magnitude can become comparable to the sum of the absolute errors for the values at the two heights. The temperature and dewpoint temperature measurements were fairly accurate but even so, in a humid climate such as Britain where vapour pressure gradients are small, errors reached 25%. The soil temperature measurement was much less accurate, due to inaccurate reference temperature measurement. The instrumentation used for the Bowen ratio technique was new and was generally the best available, though some sacrifices in accuracy were made for the sake of robustness, e.g. in the net radiometer. Using a different measurement of reference temperature could improve accuracy – the one used was the standard instrument within the datalogger and this is not good enough for this purpose. The other important factor in maximising accuracy is to ensure the gradient in air temperature and dewpoint temperature is as large as possible. The various limitations on the height of the arms have been discussed elsewhere (see Appendix D) but in terms of measurement accuracy, maximal spacing is essential.

The effect of these errors, in addition to other instrumental errors meant that if all errors in the Bowen ratio approach were at their maximum and there was no compensation, errors could reach 70%. However statistically it was found this level of error was reached less that one in 50 000 times (assuming all errors are random) and in the majority of cases the error was between 5 and 10%. This is similar to the results of Bowen ratio error analysis carried out by other authors such as Revfeim and Jordan (1976), Perrier *et al.* (1976) and Seguin *et al.* (1982) who all estimated errors at around 10%

In terms of rainfall, some accuracy was lost due to the gauge density – error would be reduced if more raingauges were installed, particularly if one were installed on the site itself. Streamflow measurement using Doppler meters however had very low instrumental error. However it is possible that in reality errors were higher than the given instrumental error if siting was imperfect, or if stream debris occasionally disturbed the ultrasonic signal.

Storage measurement inevitably has a large number of sources of error as it is very difficult to accurately measure water volume over such a large area, where water depths are variable and access limited, especially when estimating from a single gaugeboard reading. Relative errors for the gaugeboard readings used in the complete balance were 24% and 28% although together, again due to the problem of loss of significance error, their combined error was 299%. As the change in storage was less than 28% of the total storage volume it could potentially be completely accounted for by error. However, by the same count, because storage change was a small part of the overall balance its error was insignificant and the overall error in the water balance was just 14% which is fairly encouraging. The closed balance error was just 1.2%, which indicates that either error is less than expected, or that errors on each side of the water balance are compensating – possibly the more likely explanation.

Overall the low estimate of error for the measured water balance gives confidence in the water balance that has been created for Stodmarsh NNR for 2000 and 2001. The measurement techniques used were some of the more sophisticated available. Doppler meters are thought to be more accurate than the more commonly used weirs and energy balance techniques such as the Bowen ratio and the use of the Penman Monteith equation were recommended for evapotranspiration estimation by Lott and Hunt (2001).

It is difficult to compare error levels with those of other studies as they are presented in different ways. However, in general the estimated errors in this study compare favourably with those of other studies. Gehrels and Milamootti (1990), for example, quote a total inflow error of 21% and an outflow error of 30%, based on instrumental errors, which would be likely to result in a high overall value for the balance as the absolute errors corresponding to these relative errors must be added together. Other studies also have additional error caused by the problems of estimating groundwater exchanges – Owen (1995) uses specific yield of the soil as part of this and quotes errors of up to 60%. Gehrels and Milamootti (1990) state that hydraulic conductivity estimates can be no better than 50% accurate. This study, with its comparatively simple surface water measurements therefore appears to be relatively accurate.

Calculating the errors of the modelled water balance is even more critical. The resultant value for water requirements that we are aiming to calculate is essentially a residual term and therefore contains the errors of the other terms. For it to have any value at all, it is important that these errors are quantified. Inevitably the modelled water balance is going to have greater uncertainty than the measured, as the model is only based on the measured data. Accurate evapotranspiration modelling has been a difficulty in this research, despite using what is thought to be one of the most accurate techniques for evapotranspiration measurement. On many days the model was inaccurate, with up to 400% error, though the model is acceptable over months and seasons indicating that the errors are random.

Largely due to the impact of the uncertainty of the evapotranspiration estimation, the Monte Carlo analysis demonstrated a wide range of possible water requirements for each seasonal data set. When water requirements were large, 95% confidence limits were in the range of 20-40%, whereas in years where water requirements were small the uncertainty was more than 100%. This led to an uncertainty in many years as to whether water from the Lampen Stream would be required. The only year where no combination of error led to any requirement for water was 1987, the year with the highest summer rainfall. However assuming that the errors cancel out, it was found that a figure of some water requirement in around 70% of all years remained valid. It is more important however to know how much water is required and whether those requirements can be met with the available resources. The 95% confidence limits calculated from the Monte Carlo analysis of the water requirements model were used to put confidence intervals on these assessments. Confidence intervals were relatively wide in the medium values but narrower towards the extremes. The percentage of years where more than 50% of the Lampen Stream is required was originally estimated at 27%. It is possible to be confident that this value falls between 17% and 40%. Similarly 14% of years were originally estimated to exceed 100% - this may be between 8% and 22%. These are fairly wide ranges. As Lampen Stream availability was assessed on a seasonal basis the IHACRES model appears fairly accurate over this time scale and detracted little from the overall accuracy of the model.

The error analysis of the model is based on "most likely errors" which is less rigorous than the addition approach, but more useful as it is more realistic. The disadvantage of the water requirement model is it is dependent on daily data to determine when water should be brought onto and let out of the reserve. Models are less accurate on a daily basis, increasing error. A simple water balance model could be calculated on a monthly, seasonal or annual basis reducing errors within each component. The disadvantage with this approach is that error in the resultant change in storage becomes very large as it contains the errors of all the components of the water balance. Over a month, and particularly over a year, the volume of change in storage is usually small compared to the total volumes of rainfall, evapotranspiration etc. involved in the water balance. The recurring problem of loss of significance errors occurs, with the resulting massive increases in error in the residual term. Often using this approach, the entire change in storage could be accounted for by, for example, the error in rainfall estimation, even though the relative error in rainfall may be quite small (see Appendix I). Within this research the water requirements model also has the advantage of not including the errors in the outflow model, and the uncertainties associated with river inputs, which are considerable.

There were no published modelled water balance studies with error analysis with which to compare results. The significance of the above findings are that it is important to consider the confidence limits of the model when applying it and assumptions based upon it should be made with appropriate caution. When considering the climate change scenarios it must particularly be remembered that in addition to the above errors there are great uncertainties within the climate change scenarios themselves and therefore additional care must be applied when interpreting these results.

8.7 CONCLUSIONS

The instrumental errors of the complete measured water balance are 14% which is an encouraging figure and compares well to other authors. This gives confidence in the accuracy of measurements made of the water balance in 2000 and 2001. Confidence limits were also applied to the water requirement model. Uncertainty was mainly due to errors within the evapotranspiration model when applied on a daily basis and errors in water resource estimation (the IHACRES model) had little impact.

CHAPTER 9 – Conclusions and Recommendations for Further Work

This chapter aims to bring together the research presented in this thesis in order to draw some conclusions. The objectives of the work as set out in Chapter 1 are reiterated and then the extent to which these objectives have been fulfilled is discussed. This is followed by recommendations for further work, which will go further towards fulfilling the aim of the study.

9.1 MEETING THE OBJECTIVES

9.1.1 Objectives

The aim of this research was to improve the management of Stodmarsh National Nature Reserve for the benefit of the wildlife habitat and the species it supports. This was achieved by increasing understanding of the hydrology of the reserve, a key determinant in the quality of the wetland habitat. The primary objective is to calculate the amount of water that is used by Stodmarsh National Nature Reserve in a year of particular rainfall. This will enable determination of whether current water resources are likely to be sufficient in the future, and if not, the quantity and frequency with which water must be extracted from alternative water sources. Within this brief, the greatest area of uncertainty within the water balance of the reserve is the rate of evapotranspiration from reedbeds. Therefore a further objective is to increase understanding of the evapotranspiration rates of large reedbeds.

9.1.2 Objective 1: To calculate the amount of water used by Stodmarsh National Nature Reserve in a year of particular rainfall

The volume of water stored on Stodmarsh National Nature Reserve was successfully predicted using a water balance model over the summers of 2000 and 2001, when measured data were available. This was verified by the good match between storage predictions and storage measurements. When all components of the water balance were measured, inflow balanced outflow to within 1.2% and measurement errors were found to be around 14%. The water balance allowed the hydrology of Stodmarsh NNR to be characterised and confirmed that a simple water balance should be sufficient to model hydrology on the site. However over winter 2000/2001, modelling was far less successful and outflow exceeded inflow due to the addition of unmeasured fluxes of water, for example from over-bank river flooding.

The measurement years had well above average precipitation. The principal interest in the results of this objective, in terms of management, is the water requirements of the site when rainfall is low and the site is potentially lacking water. However, it is not easy to extrapolate from the wet measured years when no additional water was required, to confidently predict site hydrology and water levels in dry years. Wetland hydrological models are difficult to develop and the extreme weather conditions increased problems, particularly when calibrating rainfall-runoff models for the site. A three year study with no previous flow data is in itself a limited period in which to collect enough data to confidently calibrate and validate a model. Flooding and atypical conditions exacerbated this, at least at the outflow of the site, which is affected by on-site conditions.

These problems led to a change in the model used to meet the objective. The attempt to predict site water requirements by modelling the storage of water as a residual of the daily site water balance, over the course of a year for different meteorological conditions was abandoned as unacceptably uncertain with the data available. Instead, rainfall and evapotranspiration deficits on site and the necessity for additional water sources (in the form of the Lampen Stream or other source) to maintain the site in optimal condition were considered. Rather than calculating the water requirements under a particular set of meteorological conditions, emphasis was placed on the risk of drought recurrence, based on a thirty year historical data series.

On an annual basis, in 90% of years, even with no water from the Lampen Stream the reserve would be able to sustain itself, i.e. rainfall exceeds evapotranspiration. However, there is a seasonal rainfall-evapotranspiration imbalance resulting in loss of water from the site in summer, which is replenished by winter rainfall. A reedbed needs some surface water all year to remain in optimum condition as a habitat. In 70% of years, some water from the Lampen Stream is required to maintain optimal water levels during the summer. This does not mean that abstraction is required in 70% of years – much of the water comes onto the site naturally and can be held there using the sluices that are in place on site. In 60% of years the amount required from the stream is less than 10 000 m^3 and the maximum required in the driest year is 50 000 m^3. These volumes are more informative when related to resource availability, as in general the years when water requirement is highest will have lowest stream discharge. In 60% of years, less than 20% of stream discharge is required, 75% of years require less than 50% of the stream discharge but in 14% of years, even if all the water available in the Lampen stream over the summer months was diverted onto the reserve, this would not be sufficient to maintain the site in optimal condition.

In reality, the site managers would not be able to use the entire discharge of the stream on the reserve. Other users have abstraction licences on the Lampen Stream and are permitted to take a fixed amount, which is between 5 and 27% of the water, depending on discharge. The years with highest demand from other users are likely to be the same years as those with high demand from the reserve leading to potential resource conflict. After abstraction by other users, in 20% of years more than the entire available discharge would be required, and in around a third of years more than half the discharge would be required. In view of this, it may be necessary to allow conditions to fall below optimum in some years. Consideration must be given to how often drier than optimum conditions can be tolerated in order to calculate maximum water requirements and whether water from a source other than the Lampen Stream is required. For example, if up to 50% of Lampen Stream discharge can be taken, the site will be below optimum water levels one in three years. This may be unacceptable. In this case, for the historical data set, the water required in addition to the Lampen Stream will be in the range of 31 859 – 250 214 m^3, except in the driest year when 441 609 m^3 would be required. However if all the Lampen Stream water can be used, is one year in five below optimal water levels sustainable in terms of the site's wildlife interest? There are many management options that could be considered. Winter storage could be a feasible option as there is normally a surplus of water in winter.

It is important to consider these results in the light of the uncertainties associated with them, particularly if management decisions are to be made based upon them. All modelling contains a degree of uncertainty. Confidence limits were placed on the estimates of risk of exceeding Lampen Stream water resources, and whilst being acceptable, they show that absolute confidence must not be placed in given values. Confidence was greater at the extremes than closer to average values. 95% confidence was in the range of 6-13 percentage points each side of the original figures given.

The above analysis was carried out using historical data in order to determine drought frequencies and possible annual combinations of rainfall and runoff. However, this is a time of unprecedented climate change meaning that the last thirty years may not be representative of the next thirty years. The South-East of England is likely to be the area of the UK most severely affected by these changes. Under the medium-high UKCIP scenario Kent is predicted to have increased winter rainfall and decreased summer rainfall, increasing the annual imbalance. The perturbed IHACRES model predicted that stream discharge would be decreased all year round, indicating that the increase in rainfall in winter is not enough to offset the summer decrease in this cumulative model. The combined effect of these changes will be to increase the number of years when the Lampen Stream is insufficient to meet site water requirements. Although uncertainty is too great to put faith in figures, the general trend emphasises that in the light of climate change it may be even more pertinent to consider winter storage (which will take advantage of the increase in winter precipitation) or abstraction from another water source.

9.1.3 Objective 2: To increase understanding of evapotranspiration from large reedbeds.

The Bowen ratio energy balance approach was the primary technique used to investigate the characteristics of evapotranspiration from a large reedbed. It provides precise, high resolution data but this comes at the cost of sensitive instrumentation with demanding maintenance requirements. It proved difficult to maintain the instruments perfectly in the difficult working environment of a reedbed, especially when combined with a cool and humid climate. These problems led to data loss and some uncertainty in the results. Analysis of instrumental errors showed that theoretically they could be up to 70% but compensation and random effects are likely to constrain them to between 5 and 10% in the majority of cases.

Within this research the primary motivation for understanding reedbed evapotranspiration was to create models that could be used to predict site water requirements. The widely used approach of calculating reference evapotranspiration combined with crop coefficients was used. Crop coefficients were variable from day to day but in general they were less than one, indicating that reed evapotranspiration is less than reference evapotranspiration. This contradicts the value of 1.4 which has often been used in reedbed water requirement studies but which had little experimental basis. Further evidence that summer *Kc*s for a reedbed are between 0.7 and 0.9 was provided by the measured water balance for 2000 and 2001. When *Kc*s were calculated as the residual of the water balance, there was generally no significant difference between these and the mean results of the Bowen ratio method for the same periods. Similarly when the Bowen ratio data was used to create a water balance where all components are measured in 2001, inflows and outflows balanced to within 1.2%. This gives confidence in the results of the Bowen ratio energy balance approach.

The *Kc*s measured in this study are lower than those found in some other studies. It is only possible to hypothesise about possible reasons for this. An important possibility is related to the size of the site and the influence of edge effects in smaller sites. Large areas of homogenous vegetation such as at Stodmarsh often reduce evapotranspiration rates as the transpiration of the vegetation will raise the humidity near the surface, resulting in lower evapotranspiration and lower crop coefficients. Small reedbeds may be influenced by advection from adjacent surfaces, increasing net energy availability. This is demonstrated to some extent by the work of Fermor *et al.* (2001) who measured evapotranspiration using lysimeters on three sites. Two sites were small and newly created – at these sites *Kc*s greater than one were regularly measured, whereas at the large established site the *Kc*s were not significantly different to those measured using the Bowen ratio approach at Stodmarsh.

Reference evapotranspiration is based on a surface similar to short green grass. It is therefore an unexpected conclusion that a reedbed should have lower rates of evapotranspiration than the reference surface. The crop coefficient approach assumes a simple linear relationship with an intercept of zero between actual and reference evapotranspiration. This was not found. There are a number of potential reasons for differences in evapotranspiration from reedbeds and the reference surface. An investigation into these factors led to an increased understanding of evapotranspiration from reedbeds. The height of the reedbed means that it has a lower aerodynamic resistance. Although intuitively it would be assumed that lower resistance should increase evapotranspiration, lowering aerodynamic resistance in the Penman Monteith equation can have the effect of either increasing or decreasing evapotranspiration depending on radiation levels, vapour pressure deficit and the Bowen ratio. In many cases in the reedbed the decreased aerodynamic resistance resulted in a lower evapotranspiration estimate compared to the reference surface and a smaller *Kc* value. It also means that it has a greater capacity for the interception of rainwater and this has a significant impact on evapotranspiration. Evapotranspiration from a wet canopy is significantly greater than from a dry one, resulting in higher *Kc*s under these conditions.

The omega factor was used to determine the relative importance of aerodynamic and stomatal resistance factors. The reedbed is well coupled to the atmosphere with good rates of heat and mass transfer and therefore evapotranspiration is mainly determined by stomatal conductance. This is borne out by the significant relationships found between rates of evapotranspiration and stomatal conductance. Wetland macrophytes have often been assumed to be passive wicks to transpiration with little stomatal control. However measurements showed that this is not the case. Using a simple scaling up model the mean measured surface resistance was 98 s m^{-1}. This is greater than the value used in the reference equation, which is around 70 s m^{-1}. A gradient in stomatal resistance was found through the canopy, with lower leaves contributing less to transpiration than higher ones. Stomatal resistance was found to respond to net radiation and vapour pressure deficit. However, in general, measured stomatal resistance did not correspond well with surface resistance as defined by the Penman Monteith equation, and evapotranspiration was greater than transpiration, probably due to the influence of soil evaporation. This, in addition to other complexities, precluded modelling of reed stomatal resistance and directly parameterising the Penman Monteith equation for reeds.

Evapotranspiration from a large reedbed is extremely complicated. The height of the vegetation means that it is influenced by aerodynamic factors and yet stomatal resistances are more important. Stomatal resistance appeared to be responding to net radiation and vapour pressure but a model of stomatal resistance was insufficient to model surface resistance as defined by the Penman Monteith equation, mainly due to the influence of soil evaporation. After rainfall, evaporation of intercepted water appeared to become the dominant factor. Some progress has been made in increasing understanding of evapotranspiration from a reedbed. However a model could not be created and understanding of this complex subject is by no means complete.

9.2 RECOMMENDATIONS FOR FURTHER WORK

In order to increase understanding of the hydrology and water requirements of the reedbed at Stodmarsh National Nature Reserve, the most important extension to the work is to monitor over a longer period of time in order to include a greater range of conditions. Monitoring is particularly important over drought years, in order to assess the response of the site hydrology to dry conditions. This is especially necessary in improving the rainfall-runoff models. Ideally, model calibrations require a minimum of three years data, in addition to a significant amount of additional data (several more years) for validation and many studies use calibration and validation periods of five to ten years. Although an acceptable model was created for the inflow, more confidence could be placed in it if more data were available, particularly for validation. The outflow model should be recalibrated so that it will predict well in periods where flooding is not influential.

Another unresolved issue within the site hydrology is the magnitude of river input and particularly whether seepage occurs through the bund into the reserve. There were indications that this occurred during the winter 2000/2001 measurements, as the large unknown component may not have been entirely accounted for by over-bank flooding. However, no attempt to quantify this was attempted. As seepage through the bund is likely to be diffuse, it would be very difficult to measure hydrologically. A hydrochemical tracing technique may be preferable, for example using stable isotopes such as ^{18}O and ^{2}H to identify water sources and movement.

Logging pressure transducers could be used to give more information on water storage on the site. These are placed within tubewells and use pressure to measure the height of the water table, both above and below ground. These data give a constant record of water table height. Installed in several locations at different elevations on Stodmarsh NNR these could give a much more accurate picture of storage volume on the site and its change. Storage measurement was a large source of error in the measured water balance and this technique would give increased confidence in the accuracy of the water balance model.

Uncertainty in the rainfall data could be reduced by installing a raingauge on the site, which could be compared with data from surrounding raingauges in order to assess the accuracy of using historical data from these gauges as an indication of rainfall at Stodmarsh.

Evapotranspiration remained the greatest source of uncertainty in the final model used, despite hard work to improve accuracy. Where the investigation into evapotranspiration from reedbeds is taken next depends on the motivation for the research. The Bowen ratio approach is difficult, expensive and time consuming to operate successfully and its use probably cannot be justified in a site-wide, applied, hydrological study. To gain an understanding of how much water is being lost from the reedbed to create a site water balance, the most pragmatic approach may be to install some lysimeters. Although these are less accurate than the Bowen ratio technique and have far lower temporal resolution, they can be used with limited effort after initial installation to confirm the results collected so far. An alternative may be to measure evapotranspiration using remote sensing techniques. These do not require labour-intensive field data collection and provide data on areal distribution of evapotranspiration over large areas at very high resolution (Kite and Droogers 2000). However if the study is motivated by a desire for an understanding of the processes of evapotranspiration from reeds then research should proceed differently. This study only began to investigate some important aspects of reedbed evapotranspiration. For example, more measurements of reed stomatal conductance are required, particularly on consistently sunny, cloud-free days when the relationships between stomata and meteorological conditions would be more clearly seen. Measurements should be taken on more occasions throughout the growing season to determine the impact of changing canopy density. Separate measurements of soil evaporation are also needed. In order to understand the impact of a wet canopy, indicators of leaf surface wetness at a high temporal resolution are required.

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Appendix A – 2002 Bowen Ratio Data

A.1 INTRODUCTION

Chapter 3 describes and analyses the data collected using the Bowen ratio energy balance approach (BREB) in 2001. Because this forms a discrete journal paper, the 2002 data has not been incorporated into it but is presented separately in this appendix. This data is used in the analysis in Chapter 4 and also to validate the evapotranspiration model in Chapter 8.

A.2 METHODOLOGY

The Bowen ratio instrumentation was set up at Stodmarsh National Nature Reserve between 30th April 2002 and 16th July 2002. The location on the site was the same as in 2001 and the methodology was unchanged and is as described in Section 3.3.

A.3 RESULTS

Monthly summaries of meteorological parameters for the summer of 2001 are shown in Table A.01. June had the highest average net radiation (270 W m^{-2}) and the lowest total rainfall (27.6 mm). August had the lowest average net radiation (210 W m^{-2}) and the highest total rainfall (86.4 mm).

Table A.01 – Monthly meteorological data for Stodmarsh NNR May-July 2002

	Av. Daily radiation (Wm^{-2})	Daily temperature (°C)		Av. Daily saturation deficit (kPa)	Total rainfall (mm)	Number of rain days
		max	min			
May	238.3	16.5	8.7	0.330	58.9	18
June	237.7	20.2	10.6	0.276	42.3	13
July (1st-16th)	197.5	19.1	10.8	0.142	31.2	9

Figure A.01 shows the percentage of unusable Bowen ratio data found on a weekly basis.

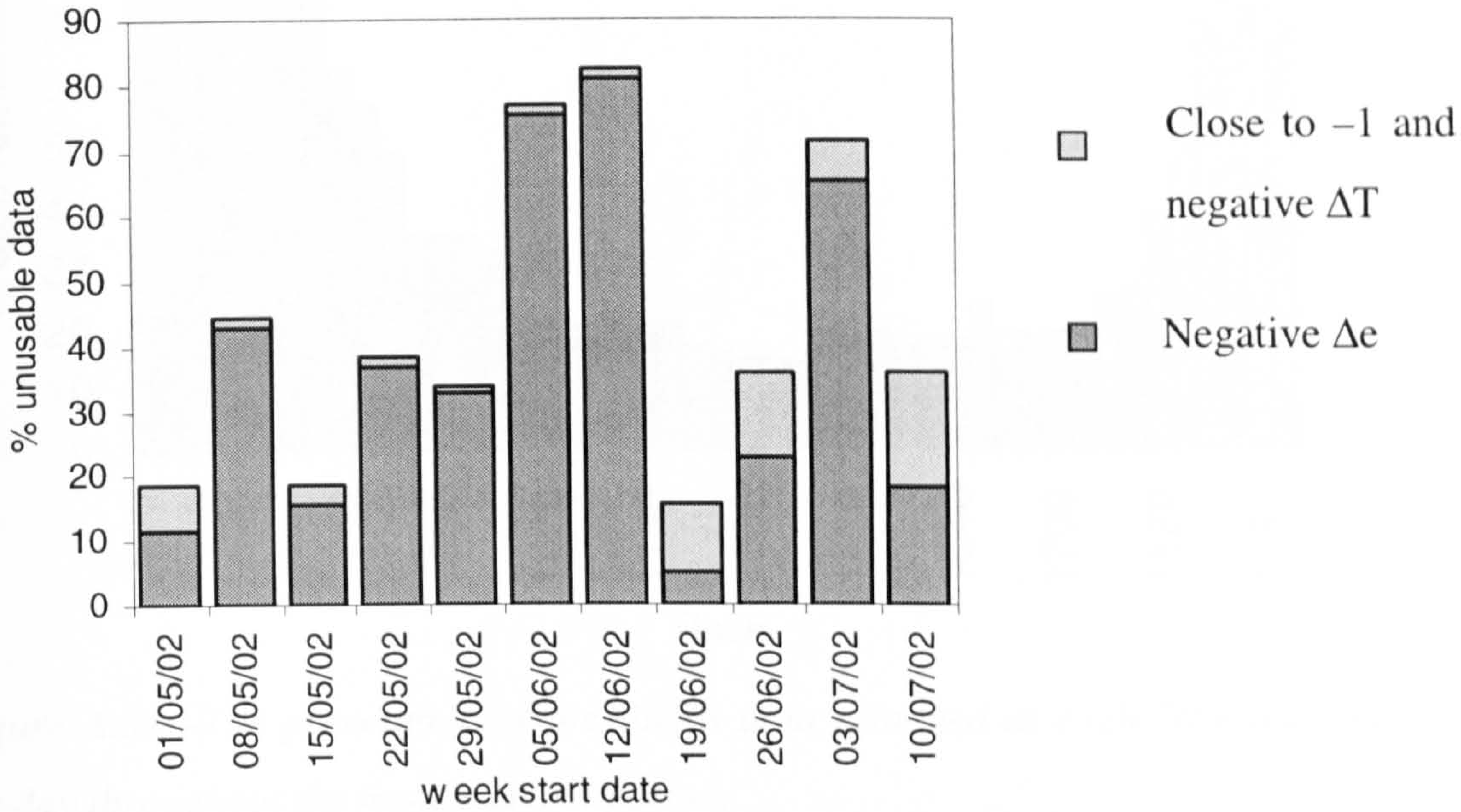

Figure A.01 – Weekly percentage of unusable data due to flux inconsistencies or the Bowen ratio being close to –1.

Generally the error ratings are much higher in 2002 than they were in 2001. In 2001 error rates were almost always less than 20% - in 2002 this is only the case in three weeks. In other weeks errors are around 40% but in three weeks the errors are up to 80%. The majority of errors are caused by negative vapour pressure measurements. The high error rates may indicate that a problem in the vapour pressure measurement had developed.

The majority of errors occurred early in the morning (Figure A.02). However there are also errors throughout the day and high error ratings in the evening, again probably due to change in flux direction.

Figure A.03 shows energy fluxes over a typical day.

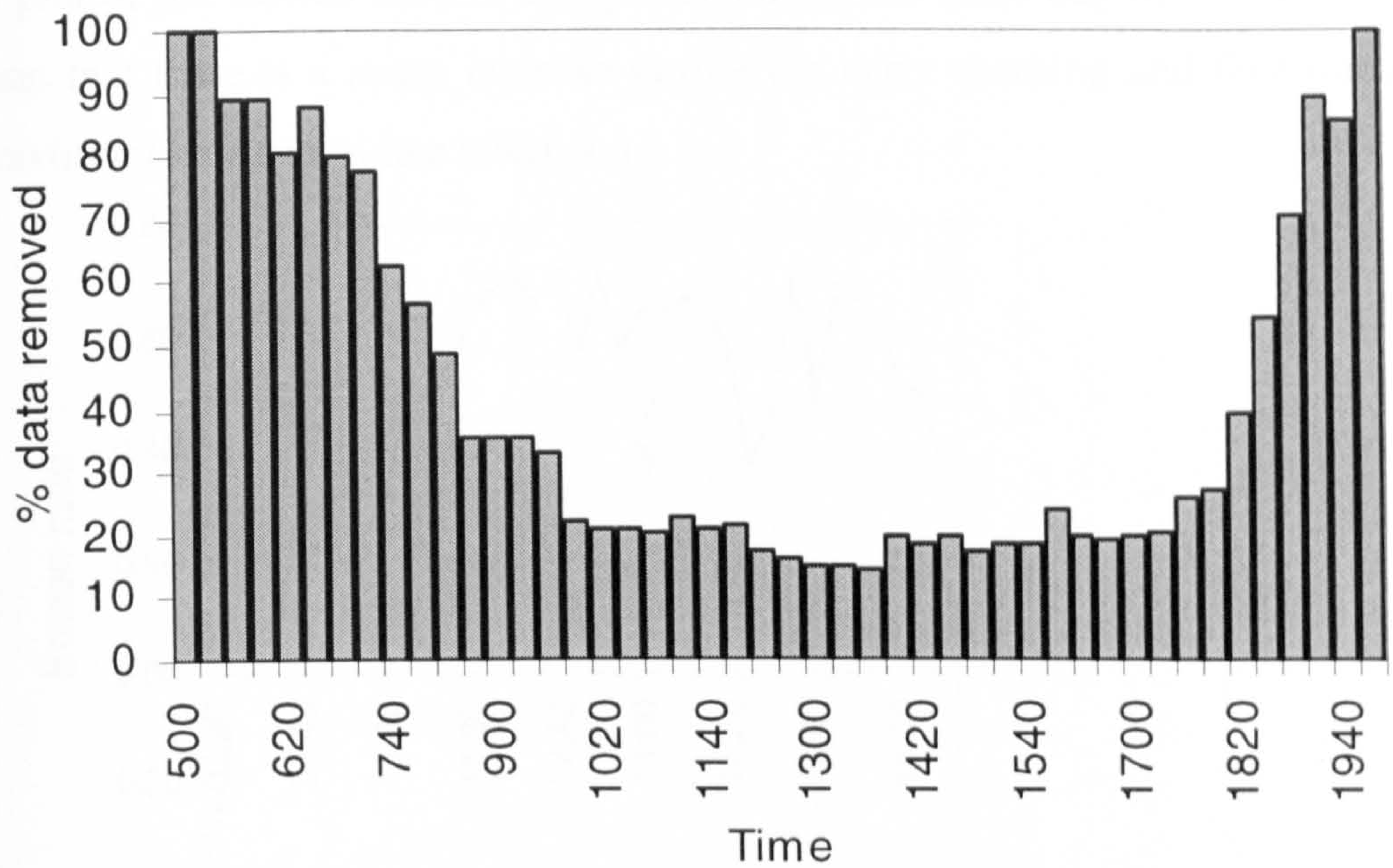

Figure A.02 –The percentage of the BREB data removed at each 20 minute period of the day throughout the measurement period.

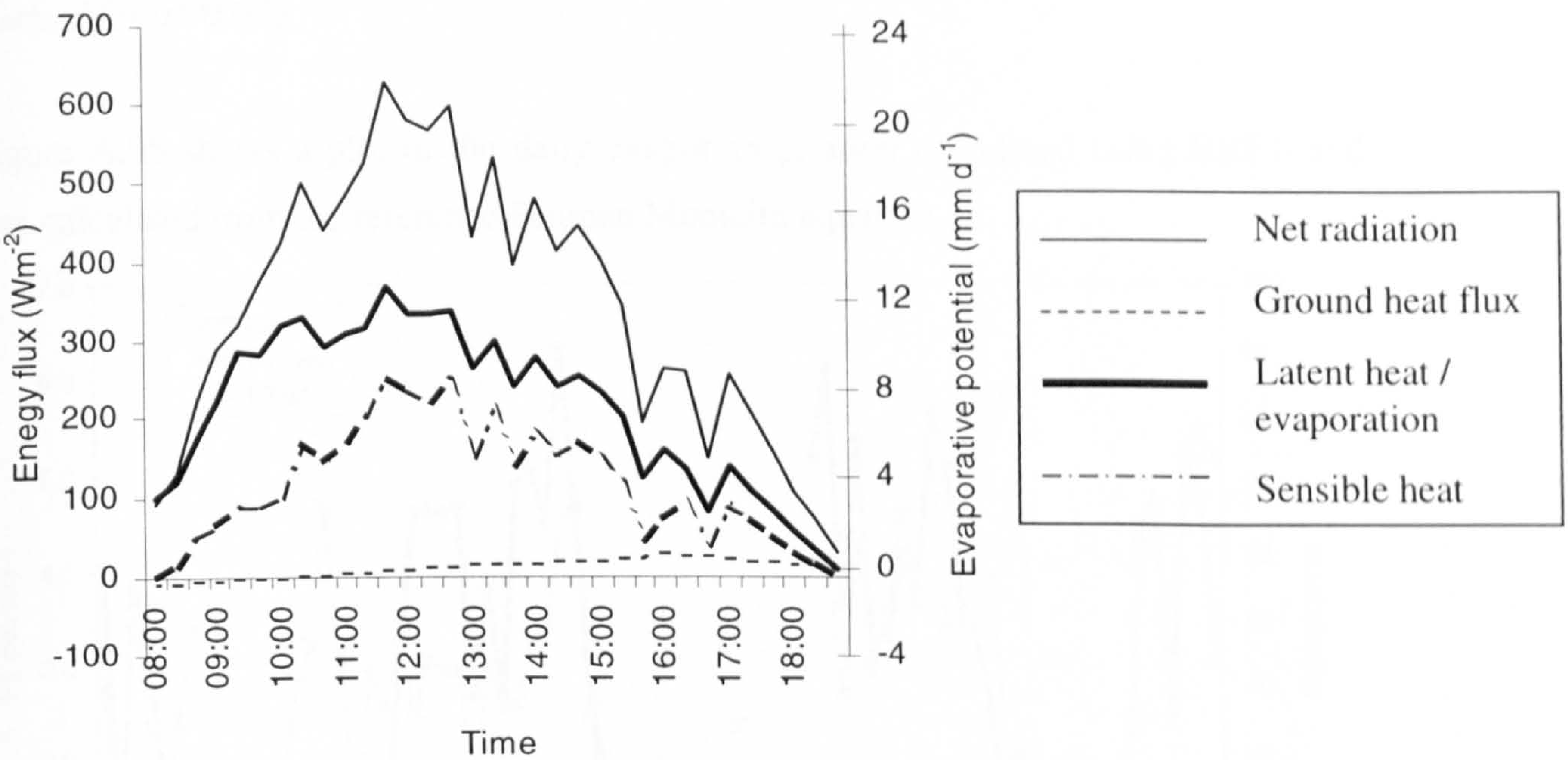

Figure A.03 – Energy fluxes for 1st May 2002 calculated using the Bowen ratio energy balance approach. The net radiation can be seen together with its division into sensible, latent and ground heat flux. All fluxes may be read on either axis – energy flux magnitude in Wm^{-2} or evaporative potential of the energy flux in $mm\ d^{-1}$.

A plot of the Bowen ratio is also presented for the same day in Figure A.04. It can be seen that there is a sharp increase during the early morning and then remains around maximum levels until late afternoon.

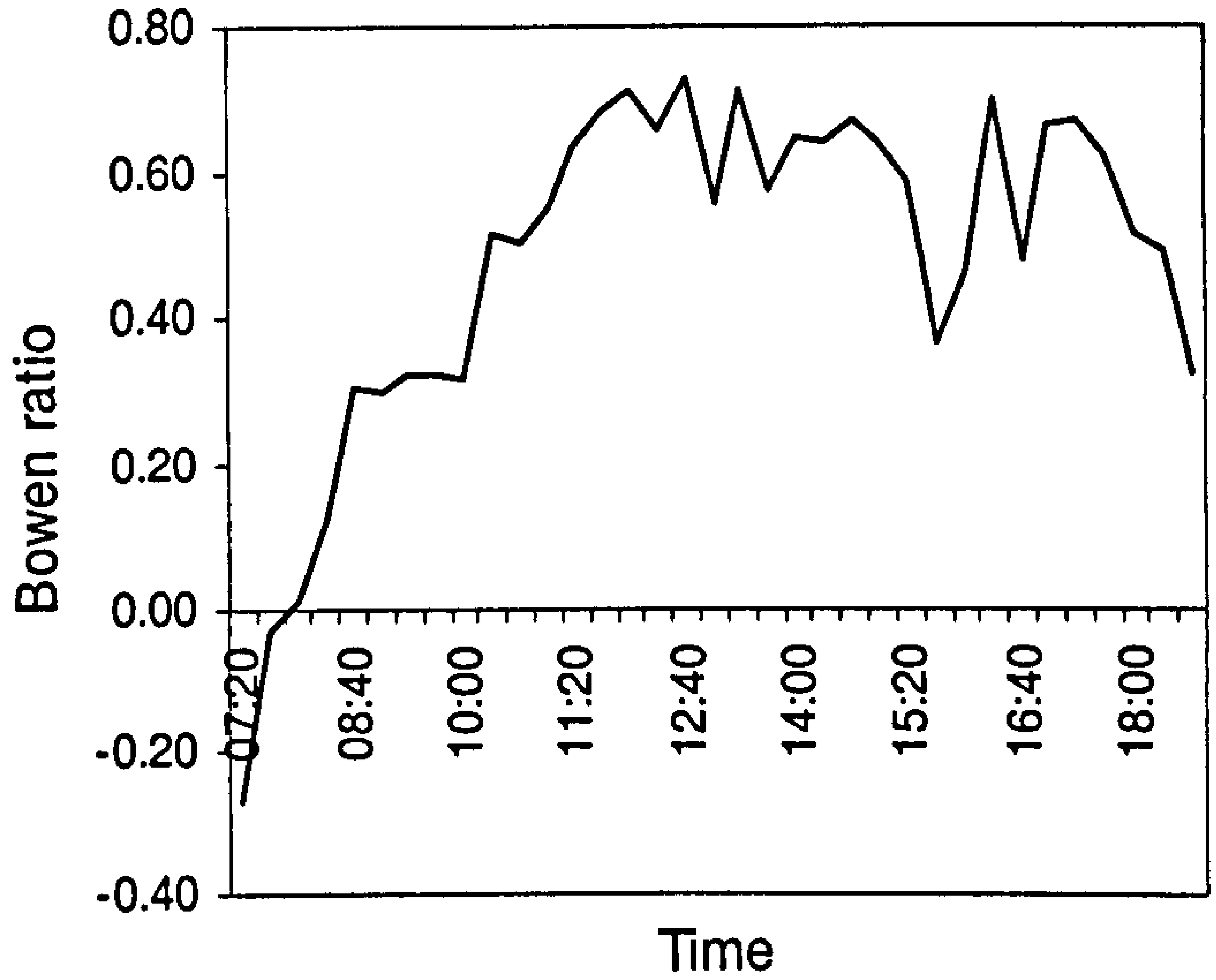

Figure A.04 – Plot of the Bowen ratio as measured by the Bowen ratio energy balance method for 01/05/02

Figure A.05 shows a plot of the daily evapotranspiration calculated using BREB and that calculated from the reference Penman Monteith equation.

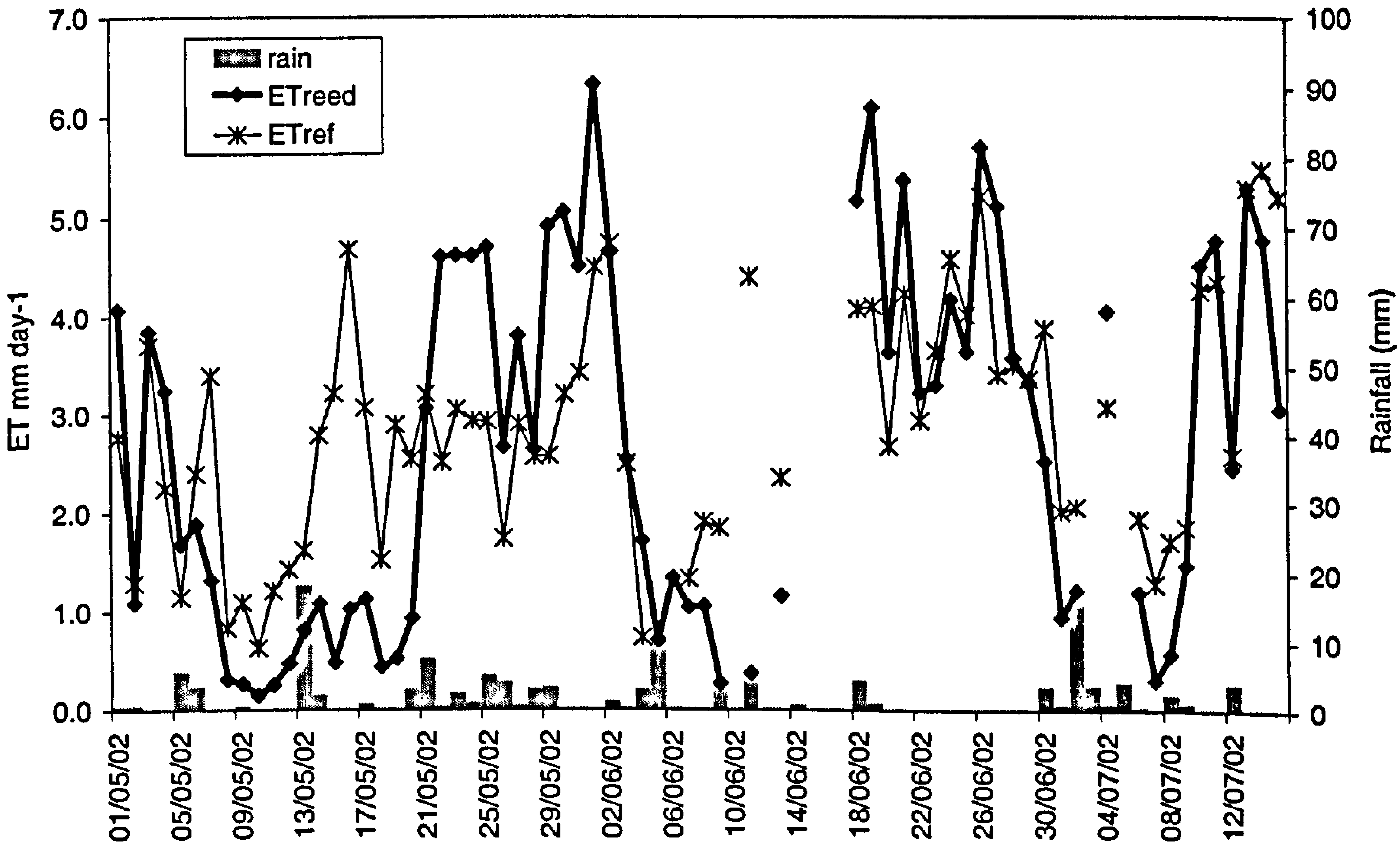

Figure A.05 – Daily average estimates of evapotranspiration from reeds (ET_{reed}) using the BREB approach and estimated from the Penman Monteith reference equation

(ET_{ref}). The gaps in the data are due to removal of data due to flux inconsistencies and instrument errors. The grey bars represent rainfall.

There appears to be little consistency in the relationship between ET_{reed} and ET_{ref}. For the first week the two data sets appear very similar, and this is followed by as period where ET_{ref} exceeds ET_{reed}. There is then a period with the opposite effect and then more days where the two data sets are fairly similar. Although there is a significant difference between days with rainfall and days without (Table A.02), it is not as strong as in 2001 and Figure A.05 does not show clear relationships between days with rainfall and days with more similarity between ET_{reed} and ET_{ref}.

Table A.02 – Comparison of Kcs between wet and dry days

	Rain	dry	*P*
Kc	1.02	0.81	0.05
n	32	33	

Daily rates of evapotranspiration (ET_{reed}) as measured by BREB plotted against calculated for reference evapotranspiration (ET_{ref}) can be seen in Figure A.06.

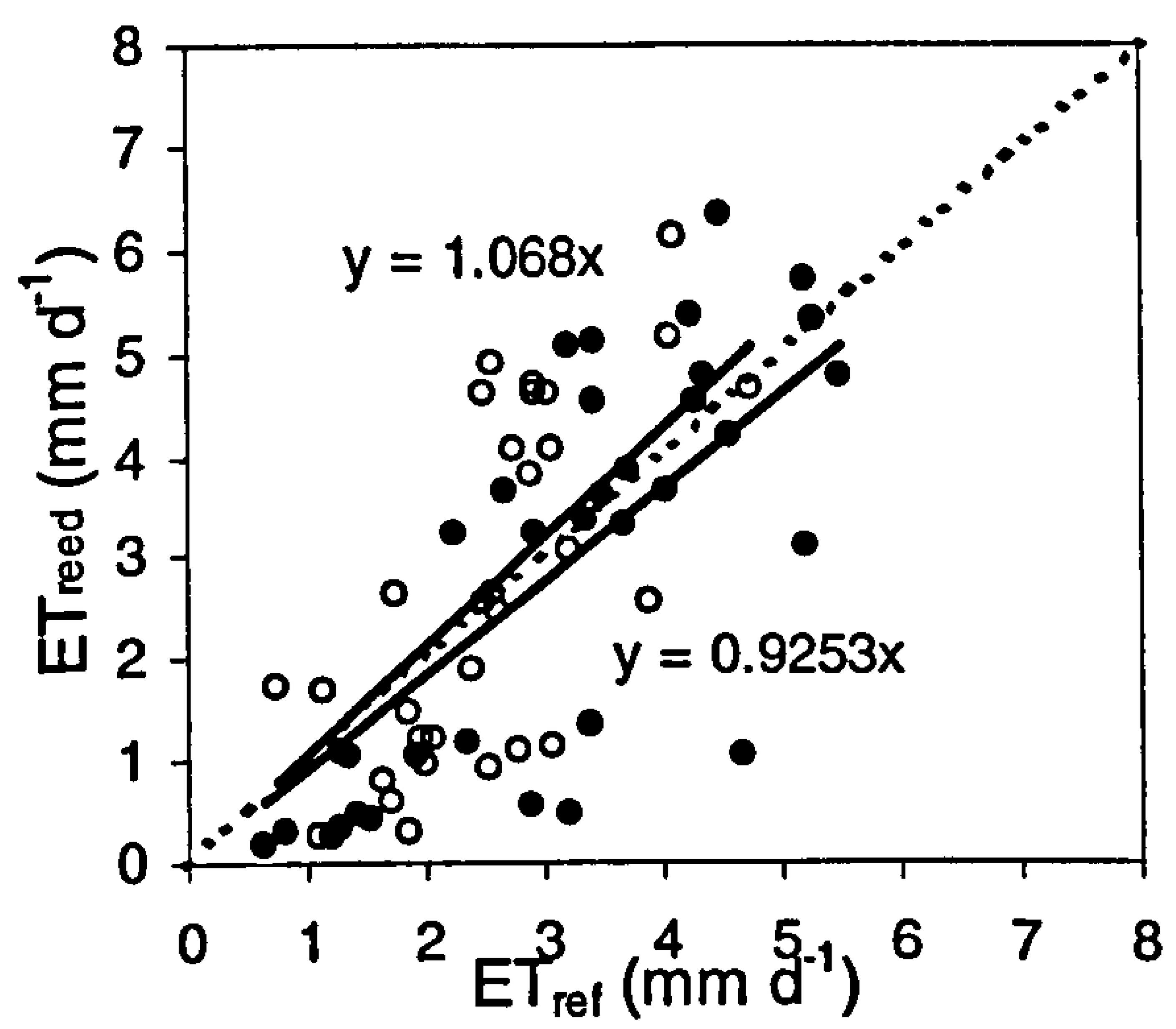

Figure A.06 - The relationship between ET_{ref} and ET_{reed} Open circles represent days with rain and closed circles represent dry days. Best fit lines forced through the origin are shown separately for wet and dry days (solid line) to create crop coefficients and the 1:1 line (dashed line) is also shown.

There is a large amount of scatter within the data and no clear relationship between ET_{ref} and ET_{reed}. There is also less clear separation of the data into days with rainfall and days without compared to 2001. Wet days have a slightly higher crop coefficient but the effect is small.

A.4 DISCUSSION

There are similarities between the BREB data from 2001 and 2002. In both data sets the range of evapotranspiration is between 1 and 4 mm d^{-1}, although there are a few extreme days measured in 2002. In both years there is no simple linear relationship between ET_{ref} and ET_{reed}. The BREB data measured in 2002 however showed a different character to that measured in 2001. One of the most noticeable differences was the smaller Bowen ratios with a greater dominance of the latent heat flux in the energy budget as compared to the sensible heat flux. The other major feature was that crop coefficients were very variable with periods where they were greater than, less than and around one, whereas in 2001 they were predominately less than or close to one. The data collection conditions were similar in 2001 and 2002. The data collection site was the same, so that conditions such as fetch and vegetation remained similar. There were some differences in the weather. Net radiation levels were lower in 2002 as were temperatures. To some extent this is to be expected as 2002 data comes from one month earlier in the year. However even the months that overlap were cooler and less sunny in 2002. Rainfall was also higher in 2002 and there were many more rain days. This increase in rainfall may partially account for the less distinct difference between rain days and non-rain days as if the climate was wetter generally the canopy may have remained wet for more days. However these differences in climate are not enough to account for all the differences and indeed the differences in different periods of data collection. To some extent instrument failure may be to blame. The dewpoint hygrometer is very sensitive and many of the errors are due to problems in measurement of vapour pressure, particularly in recording very small and negative gradients between the two arms. Many problems were experienced at the start of the research (see Appendix D) and it is possible that some of these problems may have reappeared as the instruments became older and possibly contaminated.

Appendix B – Calculating Soil Heat Capacity

B.1 THEORY

In order to calculate heat flux within the soil it is necessary to know its heat capacity. Heat capacity is an important thermal property and is influenced by the constituents of the soil in terms of the varying proportions of organic matter, minerals, water and air. The soil heat capacity (C_s) is found by adding the heat capacities of the soil constituents in 1 cm^2 of soil:

$$C_s = (C_m F_m + C_o F_o + C_w F_w + C_a F_a) \qquad \text{(De Vries 1963)} \qquad \text{(B.01)}$$

where F_m, F_o, F_w and F_a are the fractions of mineral, organic matter water and air in the soil and C_m, C_o, C_w and C_a are the heat capacitates of the same constituents. The heat capacity of air and the fraction of air in saturated soils are so small that the final term in the equation is ignored.

Heat capacities were estimated following Jensen *et al.* (1990) after De Vries (1963) as:

Minerals	1.93 kJ kg^{-1} K^{-1}
Organic matter	2.51 kJ kg^{-1} K^{-1}
Water	4.19 kJ kg^{-1} K^{-1}

B.2 METHODS

The fraction of water present in the soil was estimated using an automatic soil moisture probe (Theta Probe, Delta T Devices).

The fraction of organic matter present in the soil was estimated using the loss on ignition method. Five soil samples were taken from different parts of the reedbed. The samples were air dried for a week then the mineral and organic matter was crushed in a pestle and mortar to fit through a 2 mm sieve. Stones were removed. The samples were dried at 50°C overnight. Ten silicon dishes were weighed to an accuracy of 0.0001 g. From each soil sample two sub-samples of 5.00 g were weighed out onto the silicon dishes. These were then put in the furnace at 440°C for three hours and reweighed to determine the quantity of organic matter lost using Equation B.02

$$\% \text{ Organic matter} = \frac{(m_1 - m_2)}{(m_1 - m_d)} \times 100 \qquad \text{(B.02)}$$

where m_l is the mass of dish and dried soil, m_2 is the mass of dish and soil after ignition and m_d is the mass of dish

B.3 RESULTS

Table B.01 shows the results.

Table B.01 – Results of calculation of percentage organic mater in soil samples from Stodmarsh NNR

Sample	Silica dish (g) (m_d)	Dish + original soil (g) (m_l)	Final weight (g) (m_2)	% Organic matter
1	51.1853	56.1906	55.4066	15.66
2	37.163	42.1592	41.3522	16.15
3	49.2438	54.2448	53.6165	12.56
4	57.2292	62.2255	61.5924	12.67
5	40.6353	45.6397	44.872	15.34
6	41.8815	46.8749	46.1029	15.46
7	46.5887	51.5901	50.878	14.24
8	49.9027	54.8989	54.1835	14.32
9	45.2255	50.2283	49.5055	14.45
10	50.6108	55.6061	54.8854	14.43
Average				14.53

The soil moisture content measurements (F_w) had an average value of 0.45.

Therefore:

$F_w = 0.45$

$F_o + F_m = 0.55$

$F_m = 0.55*(1 - 0.1453) = 0.47$

$F_o = 0.55* 0.1453 = 0.08$

$$C_s = (1.93 \times 0.47) + (2.51 \times 0.08) + (4.19 \times 0.45) = 2.993 \text{ kJ kg}^{-1} \text{ K}^{-1}$$

B.4 CONCLUSION

The specific heat capacity of the soil beneath the reedbeds is around 2.993 kJ kg^{-1} K^{-1}.

Appendix C - Calculating Soil Thermal Conductivity

C.1 INTRODUCTION

Soil heat flux is an important parameter in evapotranspiration measurement when using an energy balance approach. However the measurement of soil heat flux is considered to be extremely difficult. In this study measurement has been carried out using a simple non-self calibrating heat flux plate. The use of this instrument can cause problems. Firstly the thermal properties of the soil are constantly changing due to absorption and subsequent evaporation of water. Secondly, by the constant process of wetting and drying and due to the animals living in the soil, the quality of the contact between sensor and soil is not known. However it is assumed that in saturated soil these processes will not have too large an effect. Thirdly the flow of water through the soil also represents a flow of energy, which often is misinterpreted by conventional sensors. This may be a problem in this study as there is a flow of water through the reserve. The result of all this is the quality of the data in soil heat flux measurement may be questionable.

In addition it is assumed that the thermal conductivity of the plate is similar to that of the soil. Errors increase as the thermal conductivity of the soil deviates more from that of the plate. As the soil at Stodmarsh National Nature Reserve could be considered to be non-standard due to its high water content, an investigation was carried out into the validity of the use of the heat flux plate. The heat flux plates being used are Hukseflux HFP01 with a thermal conductivity of 0.8 W mK^{-1}.

C.2 METHODS

In order to calculate soil heat flux, thermal diffusivity (the change in temperature in one second when the temperature gradient changes by 1°C cm^{-3}) was calculated. To do this, the soil temperature was measured at 0.02 m, 0.04 m and 0.06 m below the soil surface in the reedbed at Stodmarsh National Nature reserve using thermocouples. The temperature was recorded every five minutes for two separate periods of a week. This data was used to calculate thermal diffusivity and this data was combined with the soil heat capacity calculated as described in Appendix B to find the soil heat flux.

C.3 RESULTS

Figure C.01 shows an example of the thermocouple measurements. Each rise and fall of temperature represents 24 hours of measurements. Temperatures were at their minimum in the morning and peaked during the evening due to the heating of the soil by radiation during the day and cooling at night.

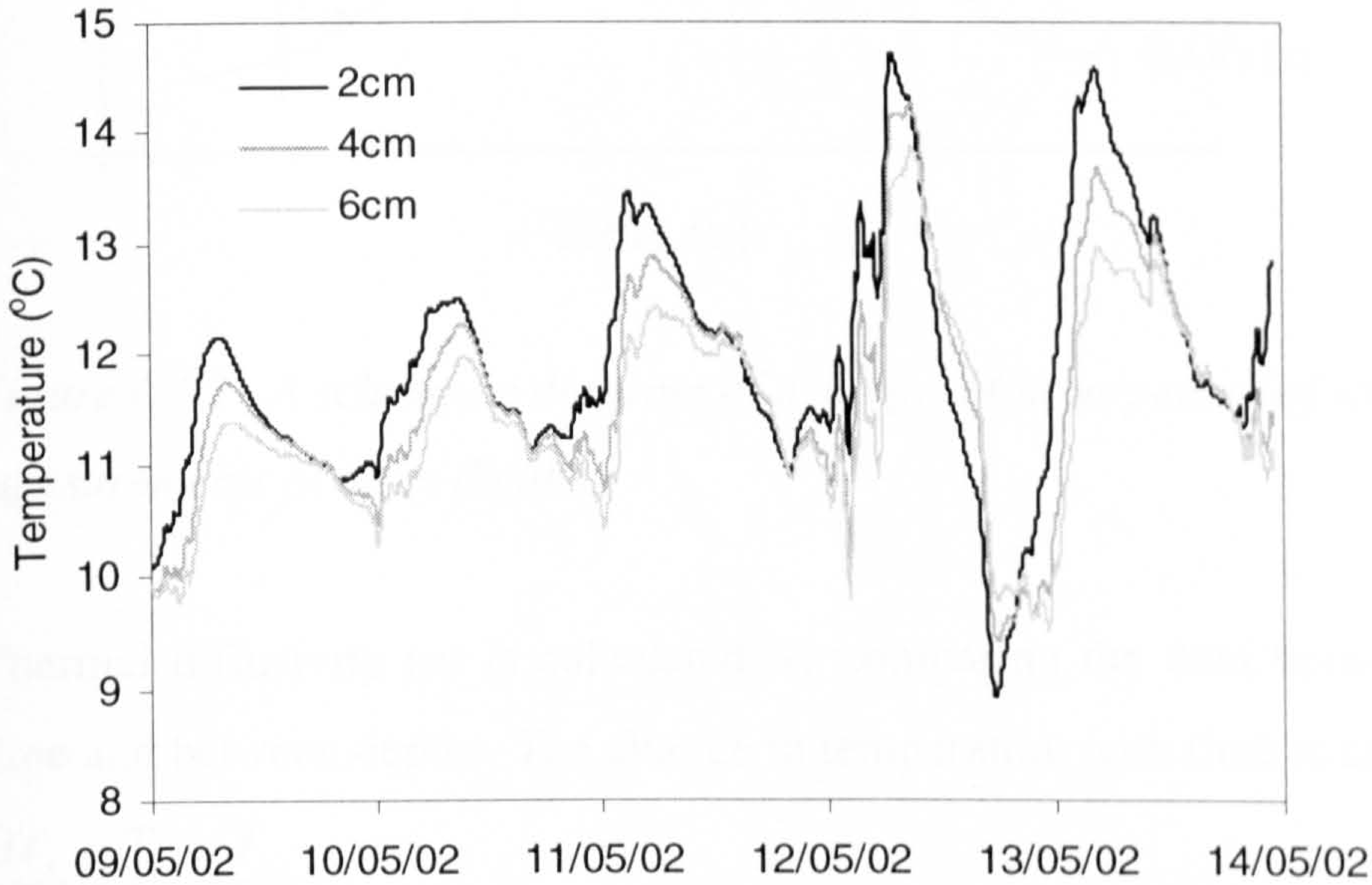

Figure C.01 – Soil temperature measurements at three depths (0.02 m, 0.04 m, 0.06 m) over five days at Stodmarsh National Nature Reserve, below the reedbed.

C.4 ANALYSIS

In order to calculate the thermal conductivity of the soil, the thermal diffusivity must be found. The data is analysed for the change in temperature with depth and the change in temperature thorough time (Figure C.02).

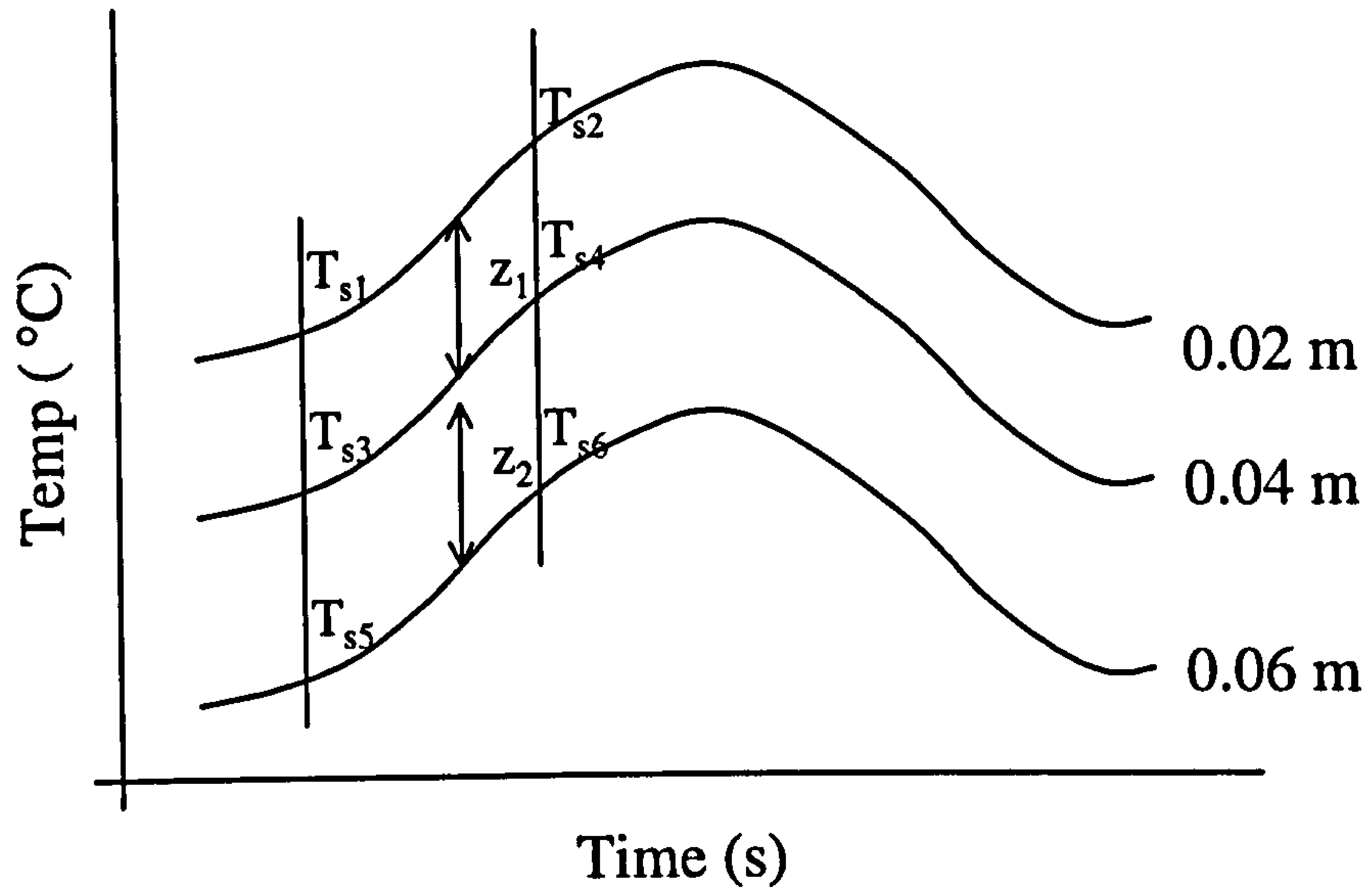

Figure C.02 - A schematic diagram of a single 24 hour period of soil temperature measurements at three depths.

Thermal diffusivity (*a*) is calculated by comparing the data between two instances in time and between depths. The change in temperature with time is calculated as:

$$\frac{\partial T_s}{\partial t} = \frac{T_{s1} - T_{s2}}{t_1 - t_2} \qquad \text{(C.01)}$$

Where T_s is soil temperature (°C) (numbers refer to positions as shown in Figure C.02) and t is time (s).This is repeated using T_{s3} and T_{s4}, and T_{s5} and T_{s6}. The mean of the three results is found. The change in temperature with depth is calculated as:

$$\text{Change in temperature between 0.02 m and 0.04 m} = \frac{T_{s1} - T_{s3}}{z_1} \qquad \text{(C.02)}$$

$$\text{Change in temperature between 0.04 m and 0.05 m} = \frac{T_{s3} - T_{s5}}{z_2} \qquad \text{(C.03)}$$

Where z is the difference in depth in mm

$$\frac{\partial^2 T_s}{\partial z^2} = \frac{\dfrac{T_{s1} - T_{s3}}{z_1} - \dfrac{T_{s3} - T_{s5}}{z_2}}{\partial z} \qquad \text{(C.04)}$$

$$a=\frac{\partial T_s/\partial t}{\partial^2 T_s/\partial z^2} \tag{C.05}$$

This analysis was carried for the days in which there was a clear rise and fall of temperature. t was set at 3600 s and analysis was done during the part of the day when temperature in the soil was rising. The analysis was repeated 210 times and resulted in a mean thermal diffusivity of the soil of 0.122 mm^2 sec^{-1} (S.E. 0.0066)

Thermal conductivity = aC_s
where C_s is the soil heat capacity (2.993 kJ kg^{-1} °C^{-1}), calculated as described in Appendix B.

Thermal conductivity = $0.122 \times 2.993 = 0.365$ W m^{-2}°C

C.5 CONCLUSIONS

The thermal conductivity of the soil is 0.365 W m^{-2}°C. The thermal conductivity of the heat flux plate is 0.8 W m^{-2}°C. The soil has a lower thermal conductivity than the plate and it is also lower than that of a typical mineral soil. However it is within the range specified by the manufacturer (0.1-1.7 W m^{-2}°C) for which the accuracy of the heat flux plate is within 20%. This indicates that the plates are acceptable for use in the soils of Stodmarsh NNR.

Appendix D - Using the Bowen Ratio Approach to Measure Evapotranspiration

D.1 INTRODUCTION

An important question to be answered within this research is the feasibility of the Bowen ratio energy balance (BREB) approach as a method of measuring evapotranspiration from reedbeds in the climate and environment of Stodmarsh. One of the reasons for the small number of studies in wetlands in general, and on reedbeds in particular is the difficulty in carrying out fieldwork. The depth of water and the density and height of vegetation restricts access. The BREB approach is complex and the majority of its use has been in arid environments over short crops. The combination of a difficult field environment with complex and bulky equipment originally designed for use in an environment which could hardly be more different is likely to make for potential difficulties in operation. These problems were compounded in this study by very limited resources available to operate and maintain the equipment.

A large amount of work was put into the task of assessing the feasibility of the Bowen ratio approach at the start of the research, both over a short grass test environment and over reeds in the field, before data collection began in June 2001. Much was learnt about what is required for the successful operation of the Bowen ratio kit. This appendix summarises some of this work and aims to give insight and help to other BREB users.

The Bowen ratio is a sensitive technique, which requires very accurate measurements of temperature and vapour pressure. These are not always easy to achieve in wet and humid climates. There is some literature on the errors involved in the Bowen ratio approach due to its inherent assumptions and published instrumental errors (e.g. Angus and Watts 1984; Bertela 1989), but little is written about problems experienced in simply getting the technique to work. The majority of authors use the same instrumentation as in the present study (Campbell Scientific Inc.). Although the Bowen ratio equipment comes as a kit from this company it was found to be insufficient to simply follow the set up instructions and leave the equipment running if a good data set was to be collected. The manual supplied by Campbell Scientific proved to be fairly inadequate in the amount of detail given in the operation of this complex equipment.

There are two types of errors that must be avoided in order to achieve an accurate data set:

- The inherent errors caused by the assumptions of the Bowen ratio approach which are not always adhered to in reality (described in Chapter 3). There are detailed papers on how to identify when these errors occur (Ohmura 1982; Bertela 1989; Perez *et al.* 1999).
- Incorrect measurements caused by inappropriate environmental or set up conditions.

The lack of published information on the latter point means that much of the following information comes from personal field experience with the equipment and from communication with other BREB users. From these contacts it became clear that it is not uncommon to have problems in getting the equipment to produce reliable data.

D.2 BOWEN RATIO METHODOLOGY

The Bowen ratio system was set up for testing over an area of short grass and subsequently in the reedbed at Stodmarsh National Nature Reserve. The system has been described in Chapter 3 and slightly more practical detail is given below which may help a potential user. A diagram of the equipment can be seen in Figure D.01.

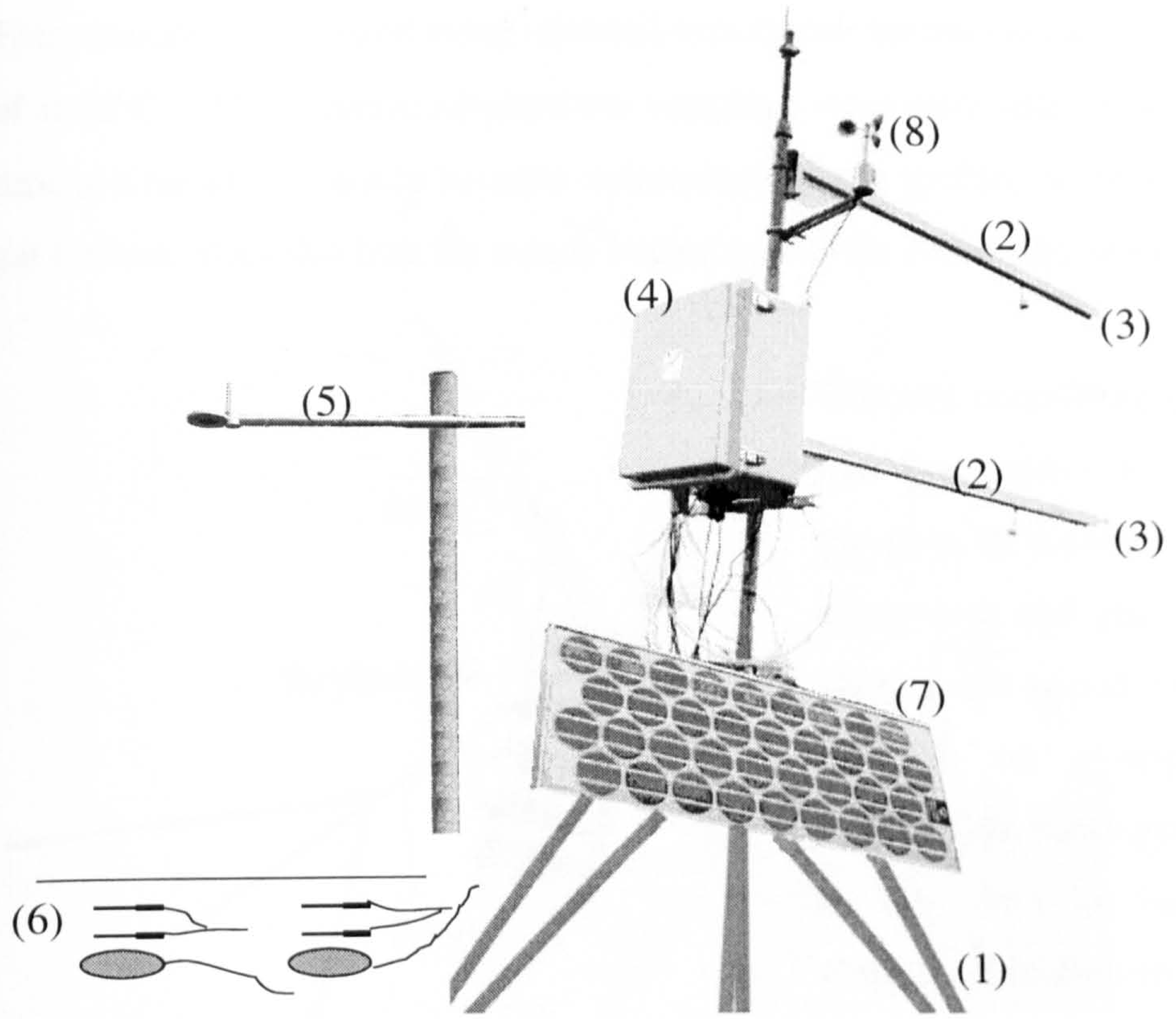

Figure D.01 – A diagram of Bowen ratio instrumentation. The numbers are refered to in the text.

The equipment is set up on a steel tripod (1). The method of setting up the tripod is described in the Campbell Scientific manual "Tripod or mast mounted automatic weather station installation manual". Vapour pressure and temperature are both measured at the end of 1.5 m arms (2) which are attached perpendicular to the tripod (they are marked as upper and lower). More detail about the height of the arms is given below. Vapour pressure is measured using a single cooled mirror dewpoint hygrometer (General Eastern Dew-10) to which air samples are sent from the two heights. The system is operated by a single low power DC pump which makes a clearly audible whirring noise when the pump is in operation. Every two minutes the air is switched from one height to the other and 40 seconds are allowed for it to stabilise. The resolution of the dewpoint measurement is 0.003°C. The dewpoint temperature is converted to vapour pressure using the equation:

$$e = 0.611 \exp\left(\frac{17.27 T_{dew}}{T_{dew} + 237.3}\right) \quad \text{(D.01)}$$

Temperature is measured using chromel-constantan thermocouples (3) with a resolution of 0.02°C. These thermocouples are supplied separately and must be changed from time to time as they easily become contaminated with spiders' webs and small insects or get broken. They slot into the purple connectors at the end of the arms.

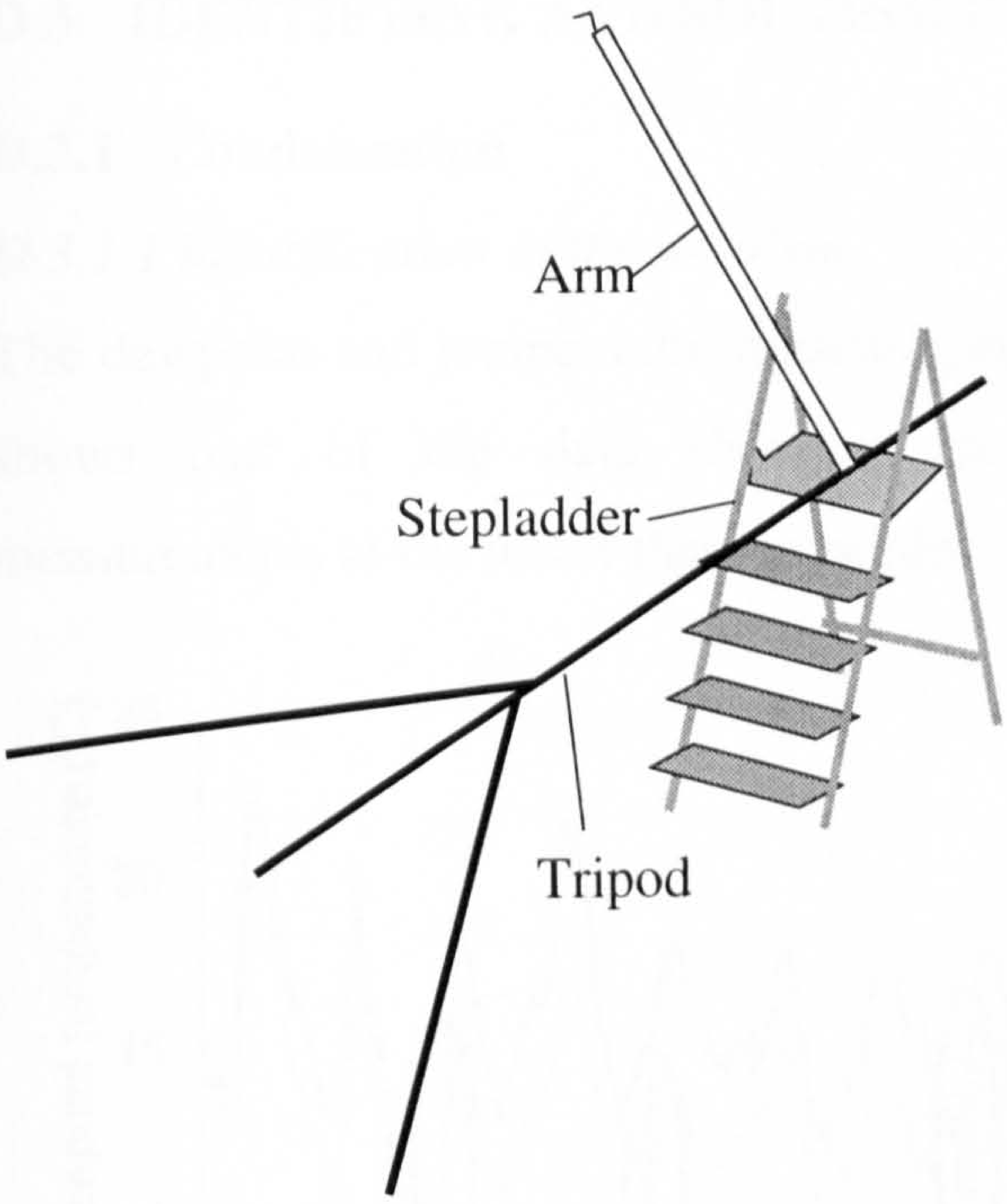

Figure D.02 - Diagram of the attachment of the arms and datalogger box to the tripod.

The box containing the datalogger (4) and hygrometer is attached between the arms. It is easiest to attach the box, upper arm and anemometer (8) is by leaning the tripod over, resting it for example on a step ladder (Figure D.02). The datalogger must be wired as described in Table 2.3-1 of the manual. The Bowen ratio programme was downloaded from the Campbell Scientific website. Some alterations were made to the program (turning off the pump at night and adding a light bulb and additional anemometers).

The anemometer is attached close to the top of the tripod. Net radiation is measured on a separate stake with a net radiometer (5) (Kipp and Zonen, NR-lite). Soil heat flux is measured with two soil heat flux plates at 0.08 m depth (Hukseflux HFP01) and soil thermocouples to measure the change in soil temperature at 0.02 m and 0.06 m depth (6). The detail of the soil heat flux methodology is given in Chapter 3.

The equipment is powered using a solar panel (7) and 70 amp hour deep cycle (caravan) battery. The solar panel is attached to the bottom of the tripod and the battery was placed in a separate battery box. The light bulb also required a separate solar panel and battery.

Maintenance was carried out as described in the Campbell Scientific manual. It was found that this had to be carried out on at least a weekly basis or error in the measurement of dewpoint temperature occurred.

D.3 IDENTIFYING AND SOLVING PROBLEMS

D.3.1 Condensation

D.3.1.1 Identification of the problem

The dewpoint and temperature measurements were analysed through time. Figure D.03 shows part of the data showing the temperature and dewpoint temperature measurements at the lower thermocouple.

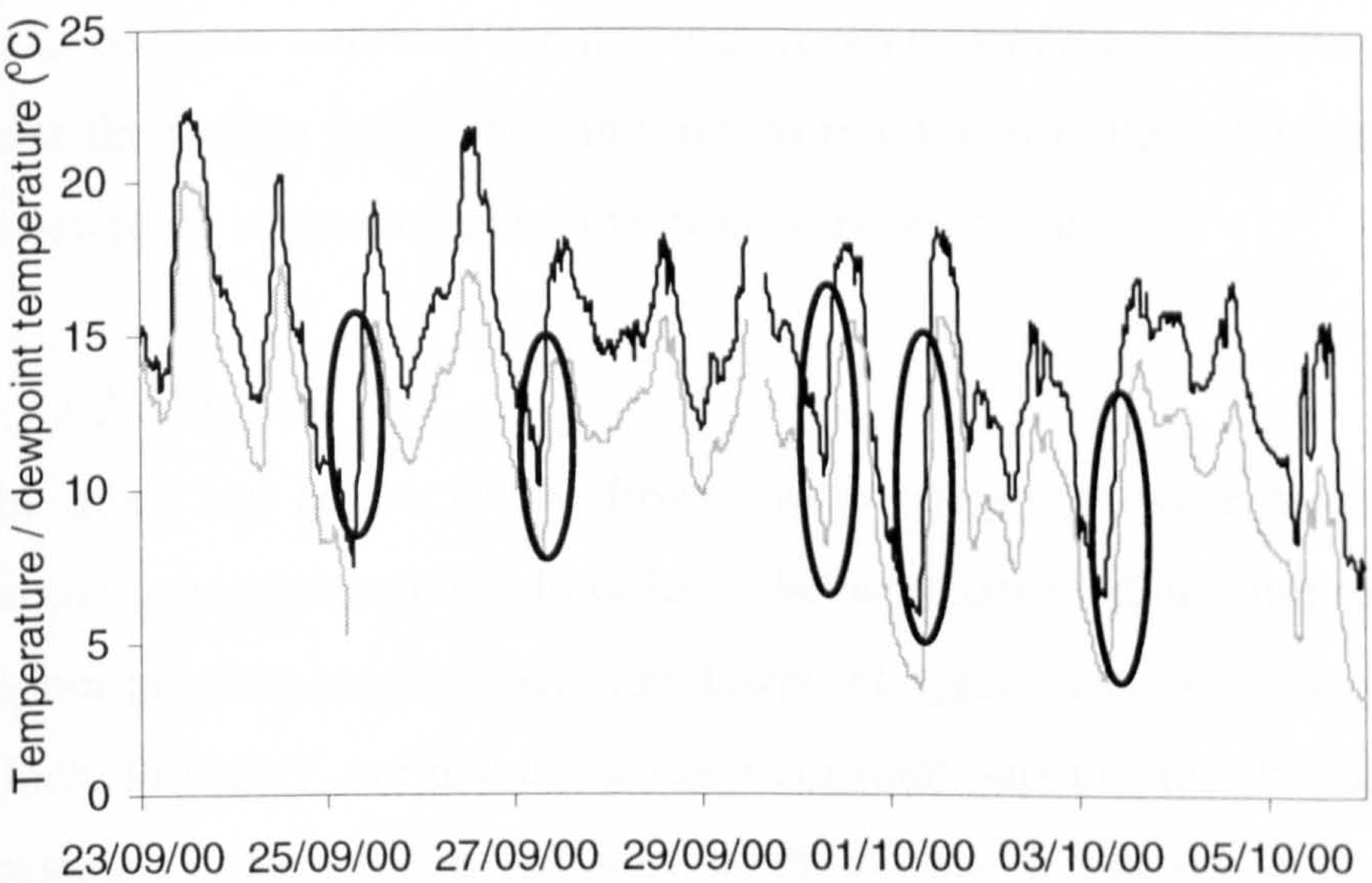

Figure D.03 – Measured temperature (black) and dewpoint temperature (grey) fluctuations. The circles represent times when condensation is thought to have occurred.

It can be seen that temperature and dewpoint temperature follow one another closely. The dewpoint temperature fluctuated more than expected. This is indicative of a problem with vapour pressure measurements. Allen (1996 p. 101) states that "any error in the [relative humidity] calibration tends to cause false variation in [dewpoint temperature] with changing air temperature. Visual observation of these false trends in dewpoint temperature can flag a problem in relative humidity readings." There were also times when dewpoint temperature appears to be greater than air temperature which is a physical impossibility and therefore indicated a problem with measurement.

The other indicator of a problem was the fact that in the morning when the temperature rises, the temperature and dewpoint temperature remain very close to one another as indicated by the circles in Figure D.03. This is indicative of condensation occurring within the hygrometer. Campbell and Williamson (1997) mention problems caused by condensation within intake hoses leading to errors in vapour pressure measurement. In a humid climate such as the UK it is important to avoid these problems. They occur because during the night the temperature may drop to dewpoint temperature causing water vapour in the air within the system to condense. As more air is drawn in, this water becomes saturated and this may continue some way into the morning until the sun heats the system sufficiently in order to re-condense the water leading to false readings of dewpoint temperature and therefore vapour pressure.

D.3.1.2 Solution

The tubes and bottles of the Bowen ratio equipment were dried out with nitrogen to remove any condensation. In order to better understand and improve the working of the vapour pressure measurement the dewpoint hygrometer was bench tested in cold rooms which, in theory, are maintained at a constant vapour pressure and temperature. It was possible to adjust the temperature of the rooms. A wet and dry bulb thermometer was used as an approximate check that the vapour pressure measurements were in the same order of magnitude and this appeared to be the case.

Campbell (personal communication 2001) suggested that to test that the dew10 is working correctly it should be turned off and allowed to rise to reference temperature. It should then be turned on again and see how long it takes to stabilise. If it stabilises within two or three fluctuations this indicates all is functioning correctly. This was done and it was found to be working correctly (Figure D.04).

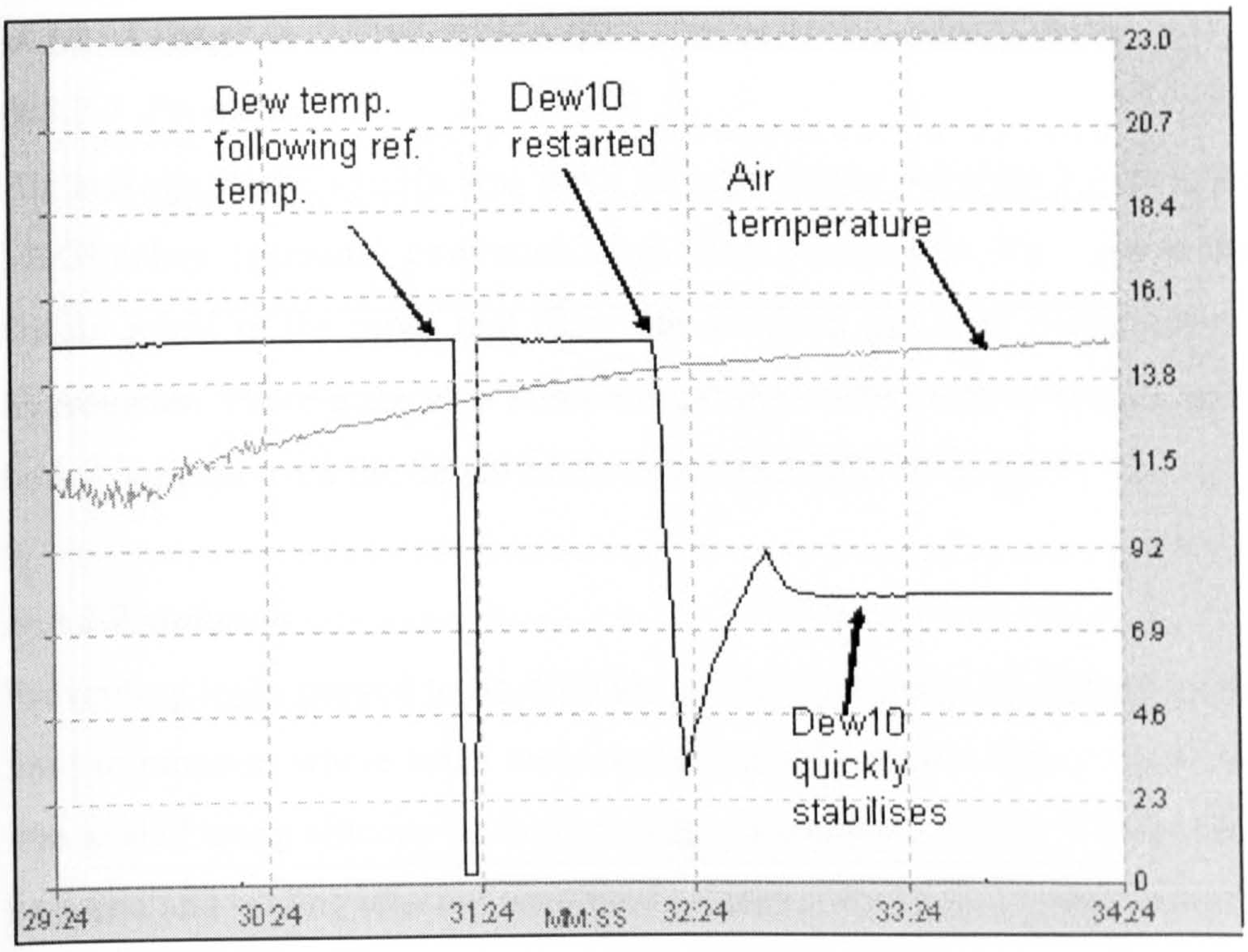

Figure D.04 - Change in dewpoint temperature (black) as it is turned off and returns to the correct value of dewpoint temperature with temperature measurements (grey).

To reduce condensation, the equipment was turned off at night. This was achieved by adjusting the Bowen ratio program so that the air pump was turned off when radiation was less than 0 Wm^{-2} and turned on again when it was greater than this figure. The other parameters were still measured at night but evapotranspiration was not calculated and assumed to be zero. A 21 watt bulb was installed in the box containing the dewpoint hygrometer in order to heat the equipment. This bulb was powered by its own battery and solar panel and was connected to the datalogger so that it was switched on at 0600 and off at 0930. The box was also insulated using polystyrene. These techniques both aimed to ensure that the temperature inside the box did not drop to dewpoint temperature therefore not allowing the water vapour in the air that was to be measured to condense.

D.3.2 Leaks

D.3.2.1 Problem

Air leakage in the system was cited as a particular problem by Rick Allen and Mark McGlinchey (personal communication 2001). Leaks in the system were discovered within joints in the pipes that draw the air from the arms where it is sampled to the hygrometer. There were also leaks around the cooled mirror block. This resulted in air being sampled from the height of the box rather than of the arms.

D.3.2.2 Solution

Preventing leaks proved to be difficult, as the equipment is difficult to get at and it was hard to pinpoint where leaks were occurring. Every joint where a leak was conceivable was sealed using silicone sealant including sealing the mirror into its block. This was a key area and sealing was repeated every time the mirror was removed for cleaning.

D.3.3 Height of Arms

D.3.3.1 Problem

A problem specific to using the Bowen ratio approach over reeds, as opposed to shorter vegetation is that there is reduced flexibility in the height of the arms. The lower arm needs to be higher than the vegetation that is being sampled so that it is sampling the bulk crop surface and not a smaller microclimate within the canopy. The top arm also needs to be low enough to ensure it is within the boundary layer of the crop, as this is an assumption of the Bowen ratio method. At the same time the arms should be as far apart as possible in order to get the best possible resolution of the temperature and vapour pressure gradient. Because of the height of the reeds, both arms have to be very high resulting in difficulties with maintenance.

D.3.3.2 Solution

In order to increase the difference between the arms, the tripod was extended in length by clamping an aluminium pole to the top using scaffold clamps. The upper arm was attached to this top pole. The length of the wires and tubes from the upper arms which connect to the box limited the height of the upper arm and therefore the distance between the arms. Some workers have set up a scaffold next to the BREB tripod in order to provide access to the upper arm for maintenance. However this has the disadvantage of causing further disturbance to the vegetation and was not tried in this study.

D.4 CONCLUSIONS

The maintenance of the Bowen ratio station in the field requires much care and the return rate of useful data can be frustrating. It is expensive in terms of maintenance. Filters must be changed regularly and thermocouples also get contaminated and must be changed. If the equipment is set up in an isolated spot as in this study, non-routine maintenance can be difficult if part of the equipment must be removed from the field site for repair. It is possible to get accurate results when measuring evapotranspiration from reedbeds though a lot of care and attention to measurement is required, both in ensuring that the instrumentation is measuring correctly and in screening the data once it has been successfully collected for inconsistencies. Simply using the equipment in a non-thoughtful way is likely to produce inaccurate results.

Appendix E – Measuring Atmospheric Stability

E.1 INTRODUCTION

In order to assess the validity of some equations used in the research, including the assessment of available fetch (Equation 3.10) and the wind profile (Equation 4.11), an assessment of the atmospheric stability conditions must be made. Atmospheric stability is the relative tendency for a parcel of air to move vertically (Oke 1993). The dominant process determining air movement in the lower atmosphere is convection. Free convection occurs due to a parcel of air being at a different density to its surroundings – if it is warmer it will rise. Forced convection occurs when vertical motion is generated by frictional interaction of the atmosphere with the surface of the earth. Often both exist together and this is described as mixed convection. Atmospheric stability conditions determine wind profiles. A true logarithmic wind profile only occurs when conditions are fully forced. This is neutral stability and occurs under cloudy and windy conditions. Buoyancy is unimportant, as clouds reduce radiative heating and cooling of the surface and winds promote mixing and mean that strong temperature stratification cannot develop. In contrast, unstable conditions occur as thermal effects become important and convection is freer – eddies become "stretched" leading to a reduction in the gradient of the wind profile. This is typical of sunny days. Stability occurs when there is an inversion, mainly at night, when air parcels are cooler than the surrounding air and sink down. This compresses eddies and steepens the wind gradient. The profiles are shown in Figure E.01.

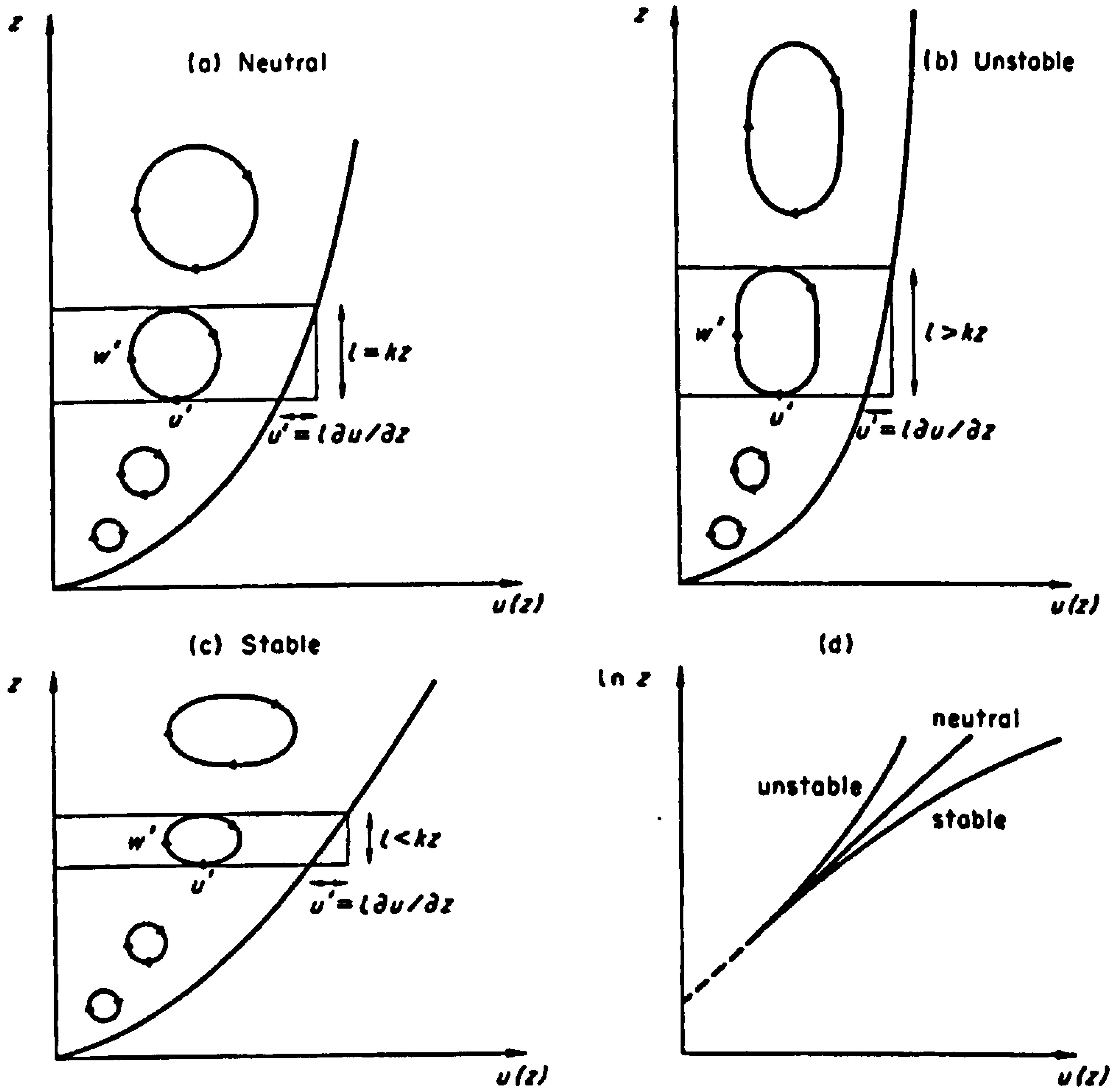

Figure E.01 - Windspeed profiles and simplified eddy structures characteristic of the three stability states in air flow near the ground (from Thom et al. 1975)

One of the best established parameters used to determine atmospheric stability is the Richardson number (*Ri*). It is the ratio of the initial turbulent kinetic energy in an air parcel to the work done on it by buoyancy forces. It is calculated as:

$$Ri = \frac{g}{\overline{T}_k} \frac{(T_2 - T_1)(z_2 - z_1)}{(u_2 - u_1)^2} \quad \text{(Thom 1975)} \qquad \text{(E.01)}$$

where g is acceleration due to gravity (m s^{-2}), T is air temperature at heights z_1 and z_2 (K), $\overline{T}_k$ is mean air temperature (K) and u is windspeed (m s^{-1}).

The interpretation of the Richardson number is given in Table E.01 below.

Table E.01 – Atmospheric turbulence conditions and associated Richardson numbers

Stability	Convection	Lapse conditions	*Ri*
Neutral	Fully forced		$-0.01 < Ri < +0.01$
Stable	Damped forced	Inversion	$+0.01 < Ri < +0.2$
Unstable	Free	Lapse	$Ri < 1$
	Mixed		$-1 < Ri < -0.01$
	None		$Ri > +2$

E.2 METHODOLOGY

Stability was assessed using the Richardson number using meteorological data measured at Stodmarsh National Nature Reserve in 2001 and used in Chapter 3 and 4 of this research. This was done using both the entire set of 5 minute windspeed data described in Chapter 4 with mean 20 minute temperature, and with mean 20 minute windspeed and temperature data for the data sets actually used to calculate evapotranspiration, after screening.

E.3 RESULTS

E.3.1 Timing of occurrence of atmospheric stability conditions

The figures below show the frequency of stable, unstable and neutral conditions over the 24 hour period using the entire set of 5 minute windspeed data.

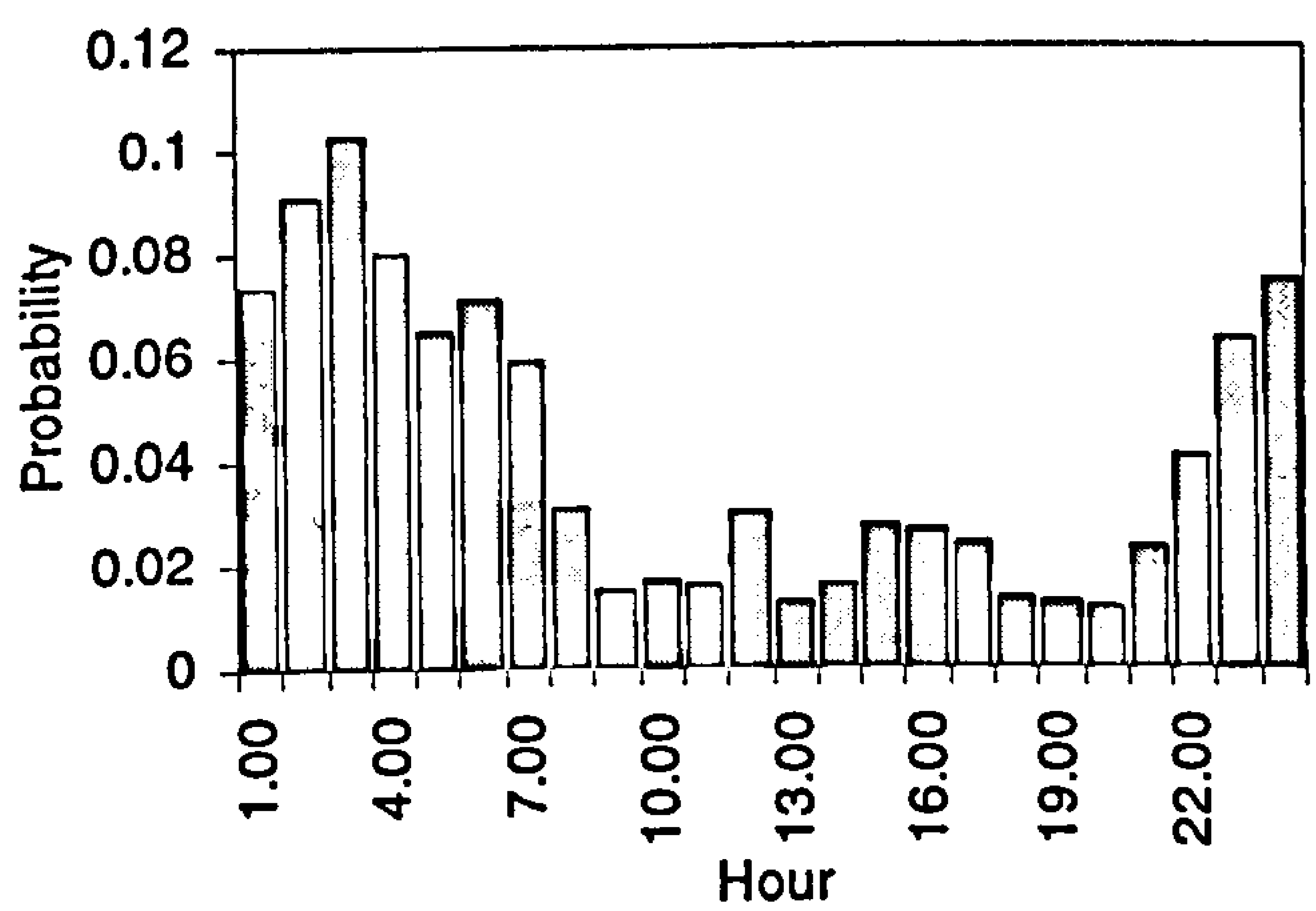

Figure E.02 – Timing of the occurrence of stable conditions.

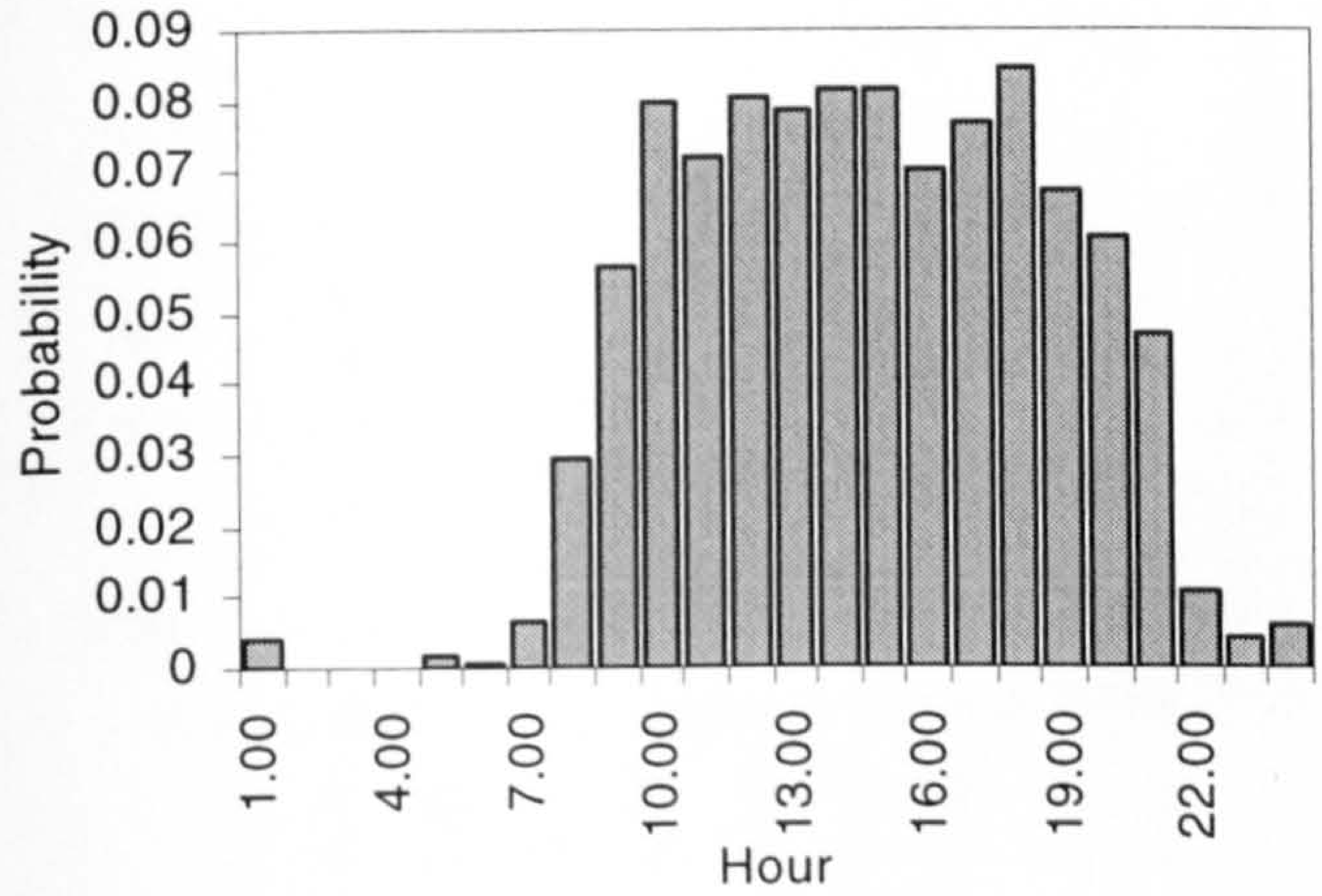

Figure E.03 – Timing of the occurrence of unstable conditions

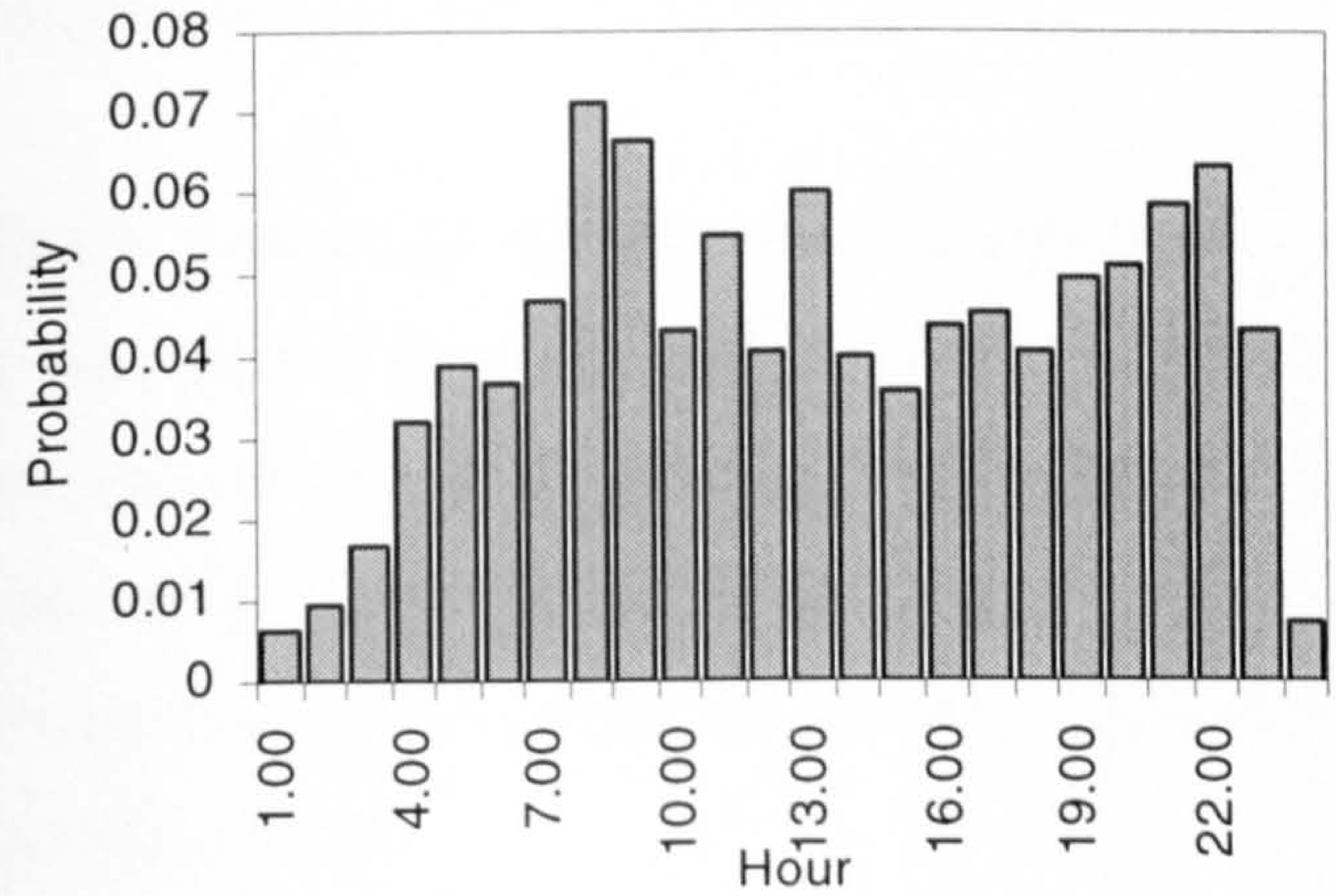

Figure E.04 – Timing of the occurrence of neutral conditions

E.3.2 Stability of Bowen ratio data

The data used in BREB were only daytime values and were screened to remove inconsistent values. Therefore the data set is slightly different to the while data set used above. Of the BREB data used to calculate evapotranspiration there were no occasions where the atmosphere was stable, 32.1 % were neutral and 67.9% were unstable.

The frequency chart below shoes the range of *Ri* values within this data set.

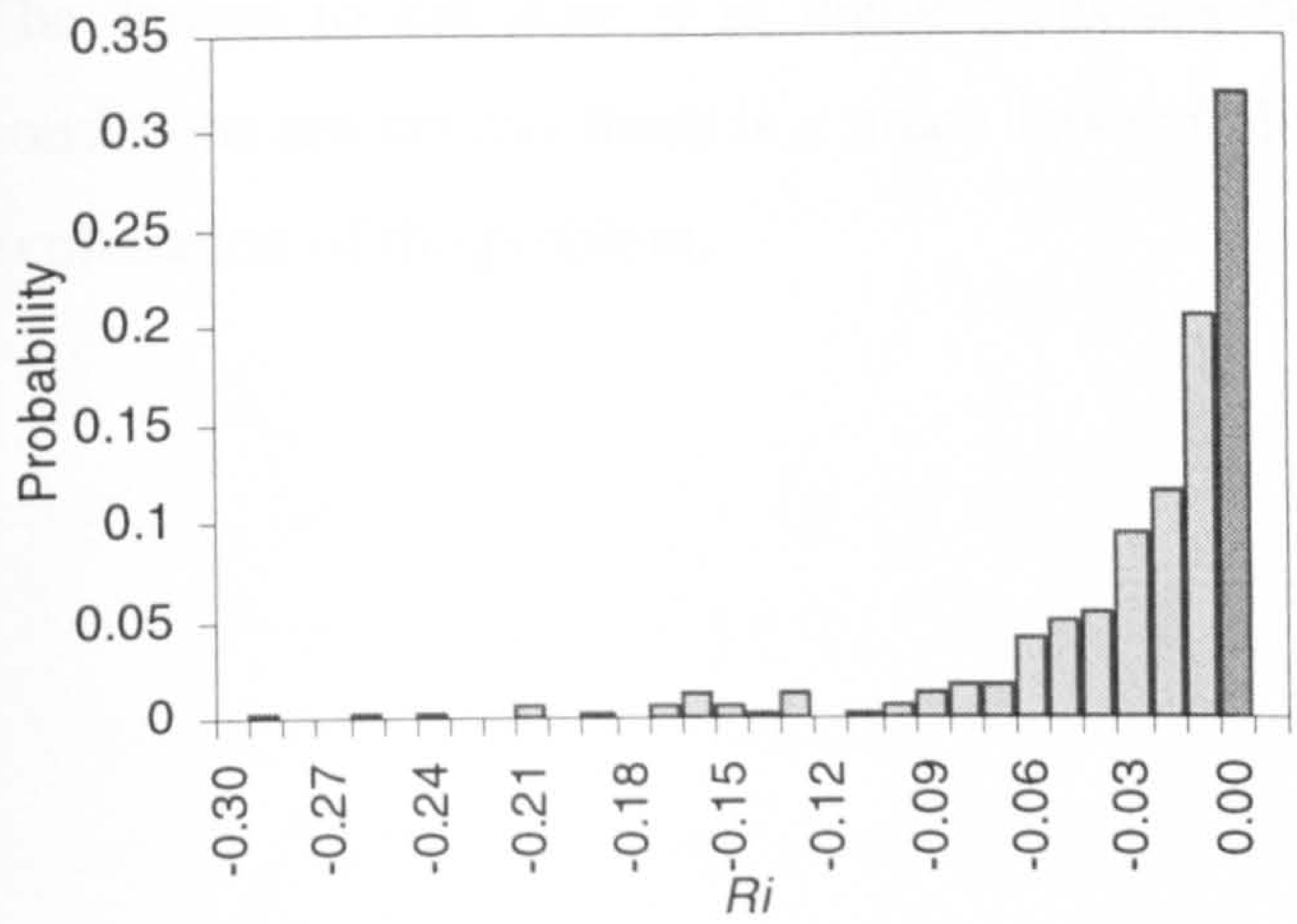

Figure E.05 – Range of Ri values within the BREB data set used to calculate evapotranspiration. The darker bar represents the values that are neutral.

E.3.3 Impact of stability on calculation of zero plane of displacement (d)

There were many data sets where it proved impossible to calculate a value of d using the iteration approach of Equation 4.12. It was hypothesised that this could be related to the prevailing stability conditions and the non-validity of the logarithmic wind profile. The frequency of unsuccessful calculation of d was calculated for each stability condition:

Stable: 63.6%

Unstable: 63.6%

Neutral: 40.8%

E.4 DISCUSSION

As would be expected, stable conditions occur mostly, though not exclusively at night when an inversion often occurs. Unstable conditions occur during the daytime and neutral conditions are also predominately in the day, though some occur at night as well. In the Bowen ratio data set however there are no stable conditions – those that occurred in the whole data set were removed in the screening process. Most of the Bowen ratio data had small values of Ri approaching neutral conditions which implies that the wind profile and fetch estimations are probably valid, or are fairly close approximations.

The failure to calculate *d* in many cases appears to be related to stability, as when conditions are neutral there is a much lower failure rate. However this is not the whole explanation of the problem.

Appendix F – Parameterising the Jarvis Model for Stomatal Resistance of Reeds

F.1 INTRODUCTION

The most commonly used and recommended stomatal resistance model in the literature is that first proposed by Jarvis (1976). Stomatal resistance is expressed by a minimum resistance multiplied by a series of functions, each representing one variable. It was originally expressed in terms of stomatal conductance but for consistency within this study it is expressed here as a resistance following Lhomme (1998). Stomatal resistance is expressed by a minimum resistance multiplied by a series of independent stress functions, each representing one variable. The general Jarvis model is:

$$r_s = r_{s\,\min} f(Rs) f(T) f(D) f(\psi) \qquad \text{(F.01)}$$

where $r_{s\min}$ is minimum stomatal resistance in optimal conditions, f is a function, Rs is solar radiation, T is air temperature, D is vapour pressure deficit and ψ is leaf water potential. The model is based on the presumption that the form of the model can be determined in the laboratory and parameters determined by fitting the model to field data using non-linear least squares regression and that that r_s responds to variables independently. The usual inputs are photon flux density at the leaf, air temperature and vapour pressure deficit above the canopy and either a set of measurements of r_s to find the parameters or a set of parameters to predict r_s (Jarvis *et al.* 1981).

Each function varies from 1 to infinity and a large number of functions have been determined for each parameter. However the model has disadvantages in that there is an assumption that there are no interactions and the parameters, once defined have little physiological meaning because the model is descriptive rather than mechanistic.

F.2 METHODOLOGY

F.3.1 Determining the functions

It is possible to determine the nature of the functional relationships between stomatal resistance and the meteorological variables by holding all other variables constant and determining the response of top down stomatal resistance solely to the change in the one variable. As top down resistance data is being used, these relationships are essentially determined by the form of the Penman Monteith equation (Figure F.01).

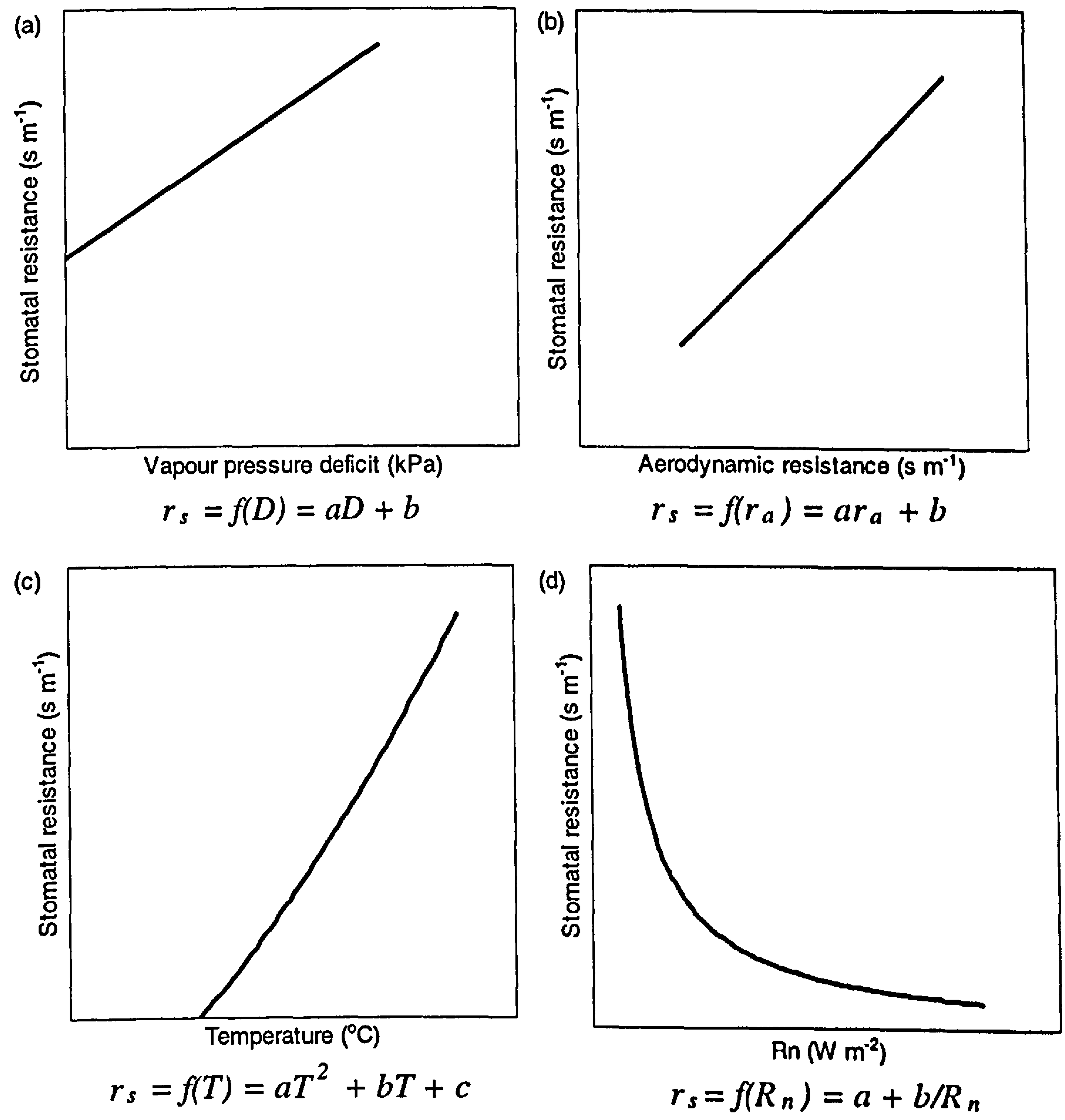

Figure F.01 – The functional relationships between stomatal resistance and (a) vapour pressure deficit, (b) aerodynamic resistance, (c) temperature and (d) net radiation determined by changing the variable in question over the observed range whilst all other variables are held constant.

This results in an overall Jarvis-type model of the form:

$$r_s = r_{s\,\min} \times \left(a + \frac{b}{R_n}\right) \times (cr_a + d) \times (e + fD) \times \left(gT^2 + hT + i\right) \qquad \text{(F.02)}$$

F.3.2 Parameterisation

With nine empirical parameters, the parameterisation of the equation is a challenge. Parameterisation was begun on a data sub-set only including occasions with >400 Wm^{-2} net radiation to reduce variation between data sets and to create a more manageable amount of data to work with. An iteration approach was used where parameters of the equation above were altered until r_s found from the equation above was equal to r_s from the top down approach. It proved impossible to fit one parameter set to the entire data set. However a set of parameters could be found for each individual 20 minute data set. The results from the subset of 73 periods are shown in Table F.01.

Table F.01 – Parameters of the Jarvis equation

	a	*b*	*c*	*d*	*e*	*f*	*g*	*h*	*i*	r_s *min*
Average	9.65	32.01	2.39	-26.36	-5.11	-70.40	34.10	-5.25	2.96	2.74
Max	26.15	32.04	14.91	4.80	1.98	2.18	89.94	-2.57	3.10	2.74
Min	-0.42	31.98	-1.69	-258.27	-345.29	-4472.54	-75.73	-11.08	2.65	2.74
S.D.	5.40	0.01	1.99	28.45	41.83	533.05	31.06	1.54	0.08	0.00

$r_{s\ min}$ was set to 2.74 and this remained unchanged by the iteration. Parameter *b* and to a lesser extent parameter *i* also showed little variation. There was large variation within all other parameters. The sensitivity to each parameter was tested. This showed low sensitivity to the parameters related to radiation (*a* and *b*) moderate sensitivity to those related to temperature (*g*, *h* and *i*) and vapour pressure deficit (*e* and *f*) but extremely high sensitivity to aerodynamic resistance, with only a very small change in these parameters causing a huge change in stomatal resistance. Because of this high sensitivity to aerodynamic resistance this parameters was then removed from the equation so it became:

$$r_s = r_{s\min} \times \left(a + \frac{b}{R_n} \right) \times \left(eT^2 + fT + g \right) \times \left(h + jD \right) \qquad \text{(F.03)}$$

It was still impossible to find overall parameters using iteration but again the parameters for each data set were fitted. However this resulted in more of the parameters becoming close to constant (Table F.02).

Table F.02 – Parameters of the refined Jarvis equation

	a	*b*	*e*	*f*	*g*	*h*	*i*	r_s *min*
Average	5.57	32.00	-0.54	0.61	51.97	-4.39	3.00	2.74
Max	5.57	32.00	-0.44	0.68	51.97	-4.39	3.00	2.74
Min	5.57	32.00	-0.63	0.51	51.97	-4.39	3.00	2.74
S.D	0.001	0.000	0.052	0.042	0.000	0.000	0.000	0.000

The functions of temperature, radiation and vapour pressure deficit were regressed against the values of temperature, radiation and vapour pressure deficit (Figures F.02 - F.04).

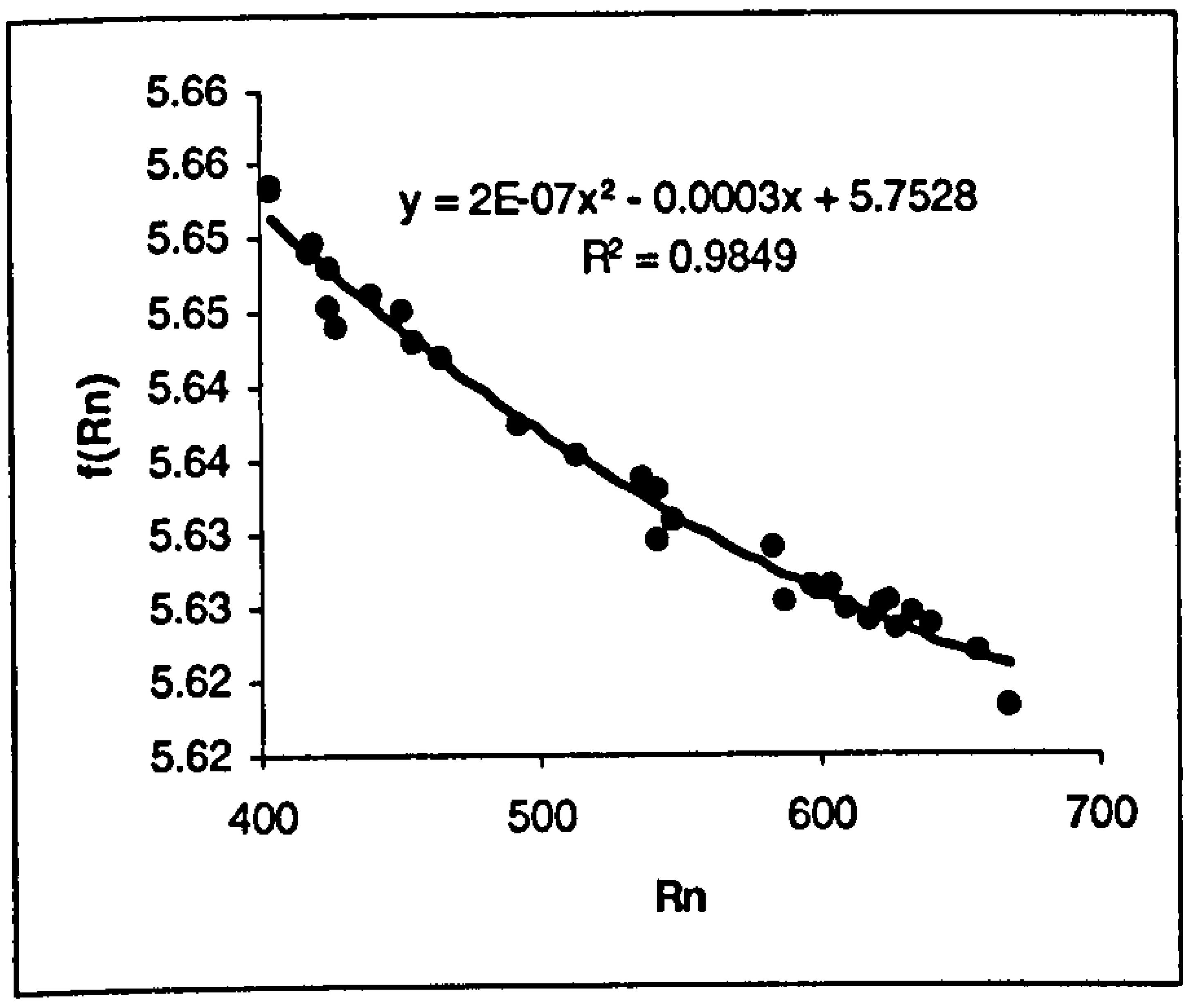

Figure F.02 – Variation of $f(R_n)$ with net radiation

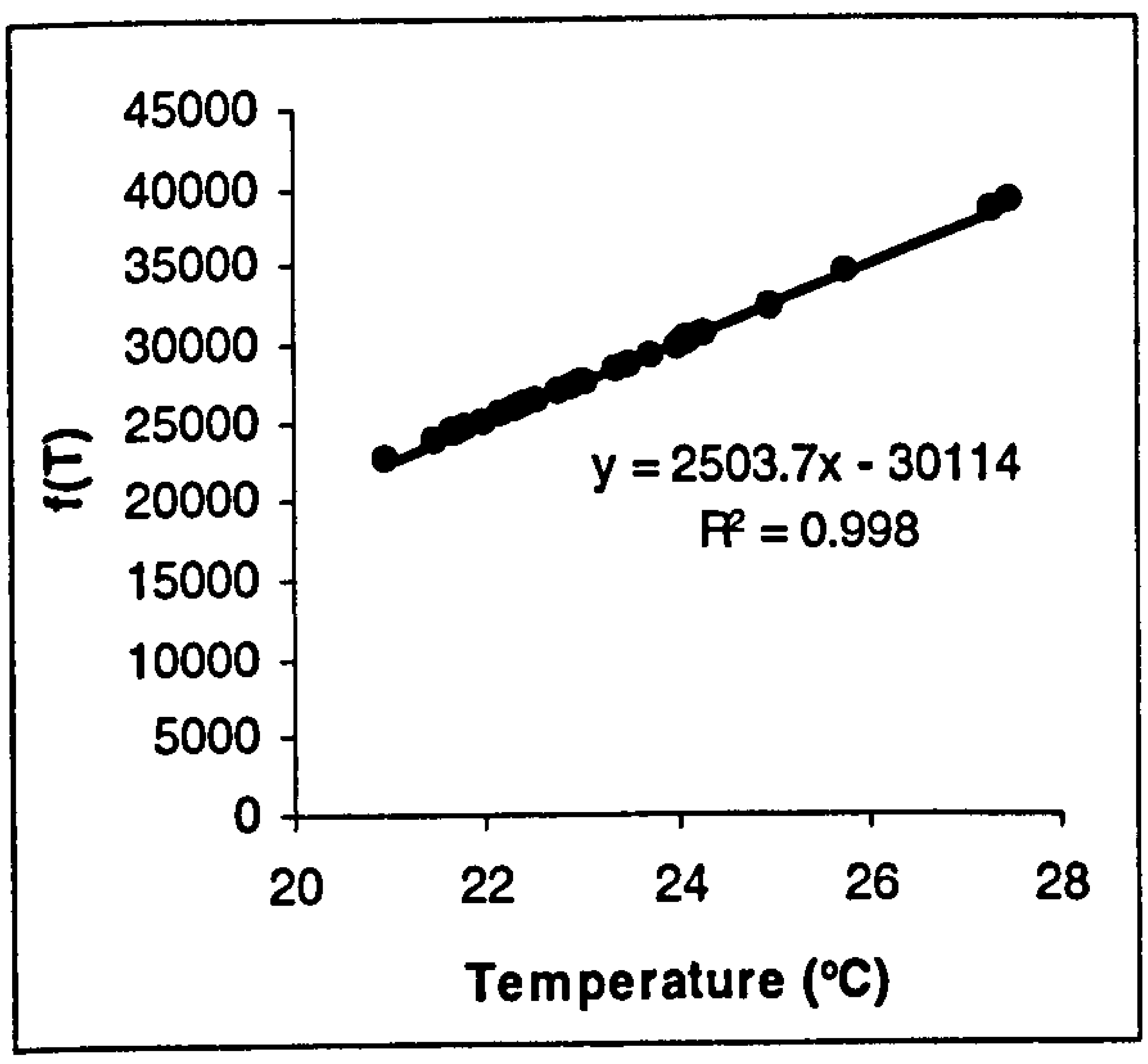

Figure F.03 – Variation of $f(T)$ with temperature

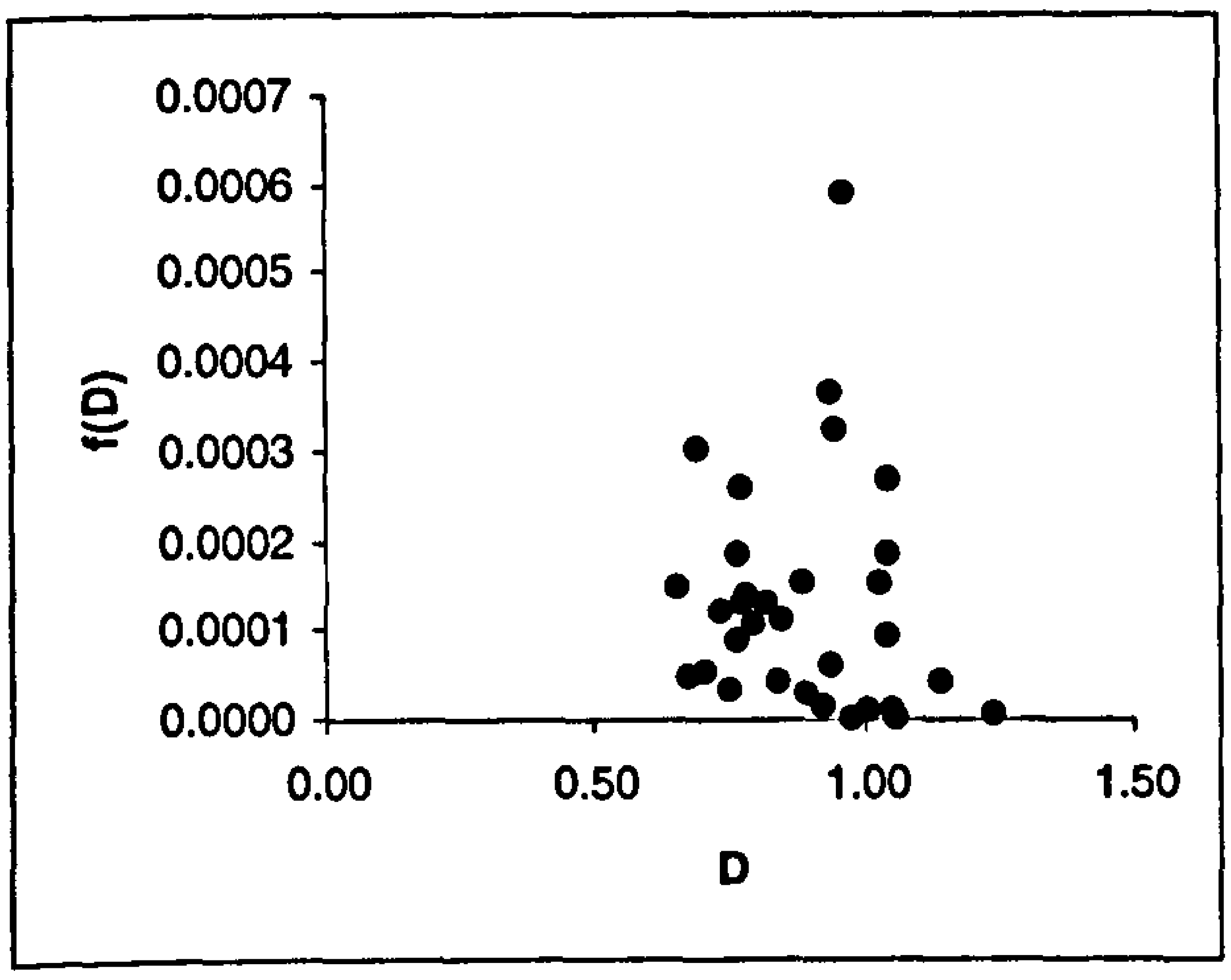

Figure F.04 – Variation in f(D) with vapour pressure deficit

Due to the consistency of the parameters of radiation and temperature there is a good relationship between their functions and data values, meaning that the functions can be calculated. There is no relationship between vapour pressure deficit and *f(D)*. However there are good relationships between vapour pressure deficit and the parameters making up *f(D)* (Figures F.05 and F.06).

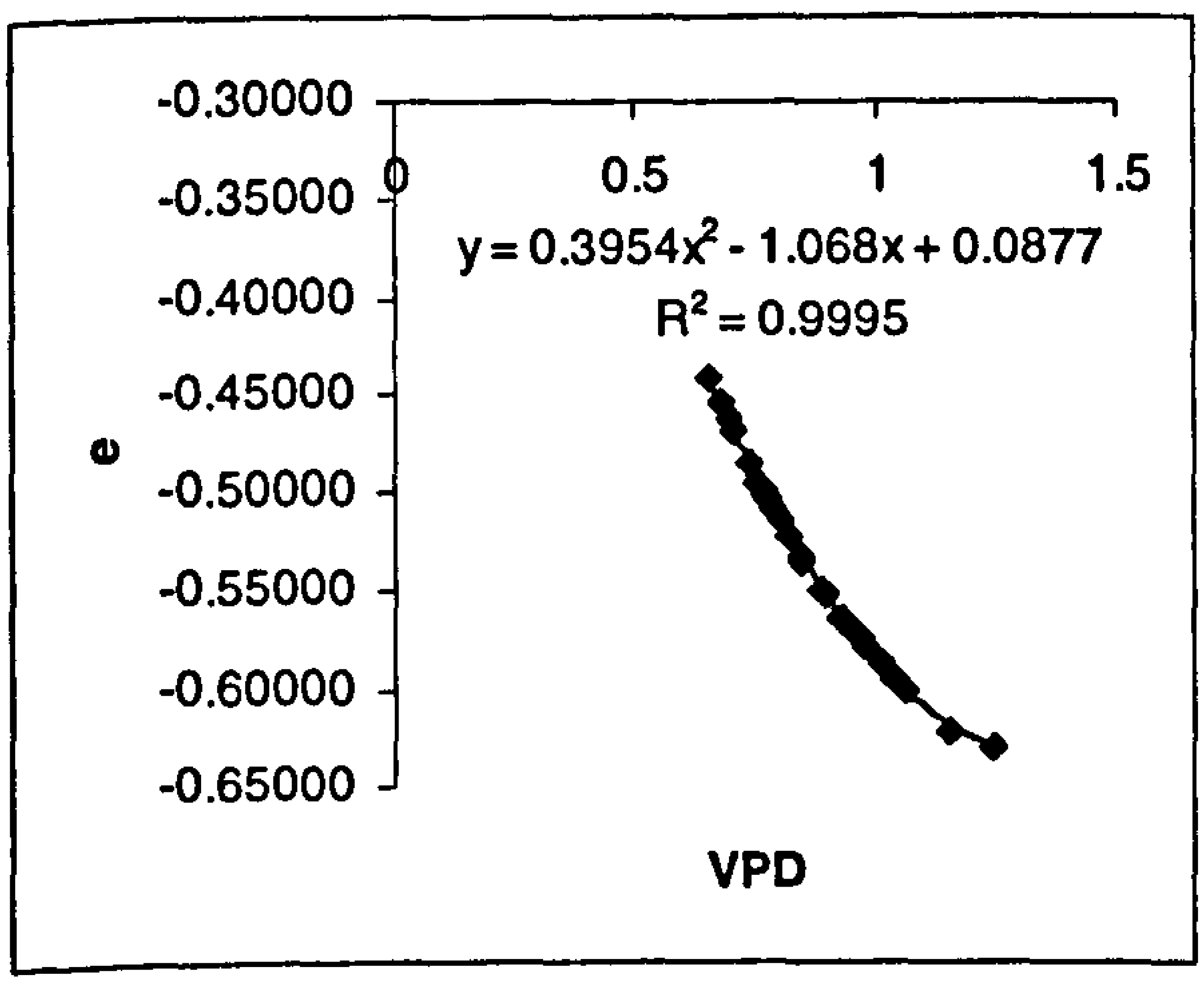

Figure F.05 – The relationship between parameter e and vapour pressure deficit

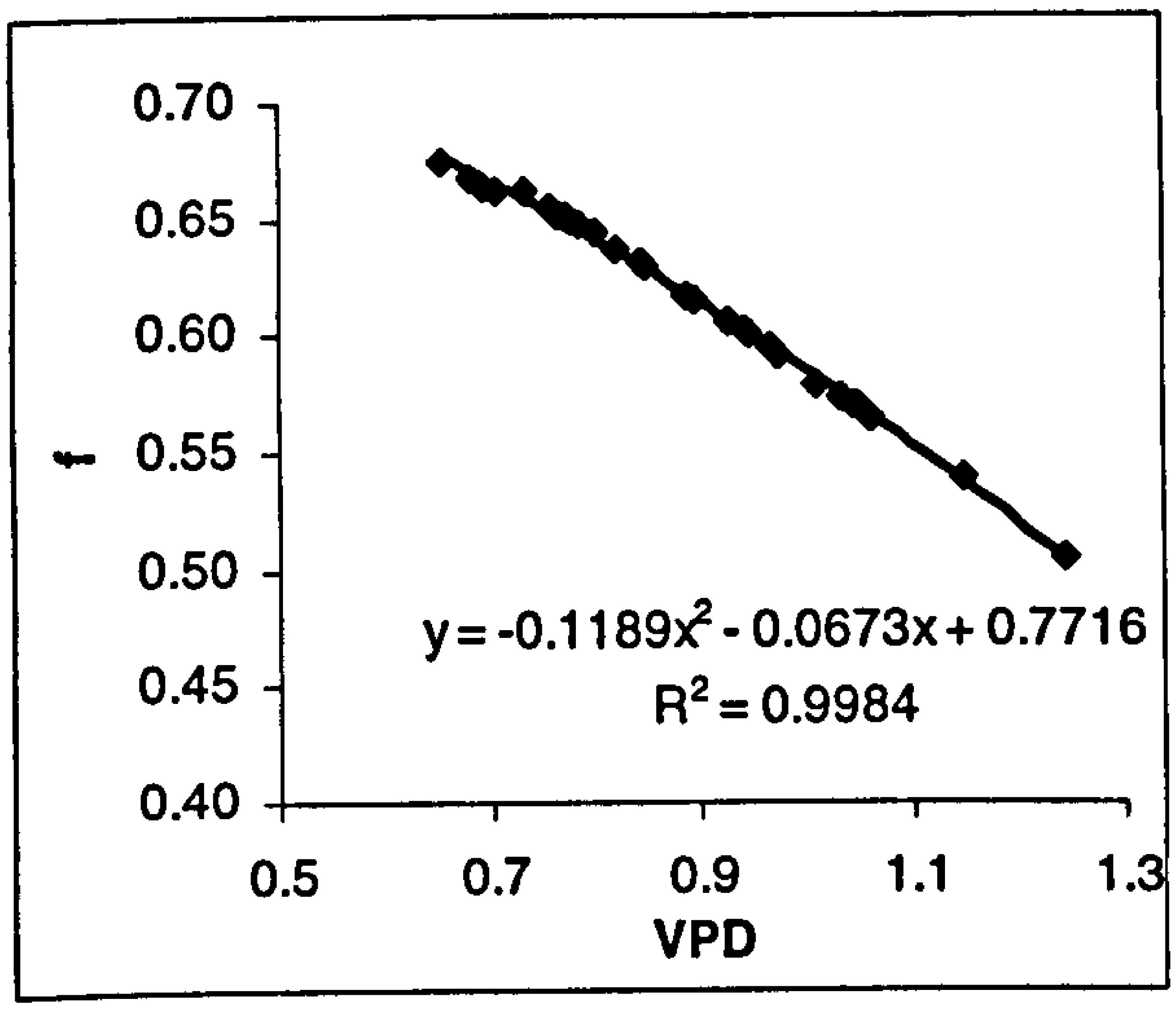

Figure F.06 – The relationship between parameter f and vapour pressure deficit

There are very good relationships to which quadratic equations can be fitted. It should then be possible to predict stomatal conductance from:

$$rs = rs_{\min} \times f(Rn) \times f(T) \times f(D) \qquad \text{(F.04)}$$

where:

$rs_{min} = 2.74 \text{ s m}^{-1}$ (values from Table F.02) (F.05)

$$f(Rn) = 5.75 + \frac{32.00}{Rn} \qquad \text{(F.06)}$$

$$f(T) = 52.00T^2 - 4.39T + 3.00 \qquad \text{(F.07)}$$

$$f(D) = e + (f \times D) \qquad \text{(F.08)}$$

$$e = 0.03954D^2 - 1.068D + 0.877 \qquad \text{(F.09)}$$

$$f = -0.1189D^2 - 0.0673D + 0.7716 \qquad \text{(F.10)}$$

F.3 RESULTS

This methodology was tried on the same subset of data used above and the results can be seen in Figure F.07.

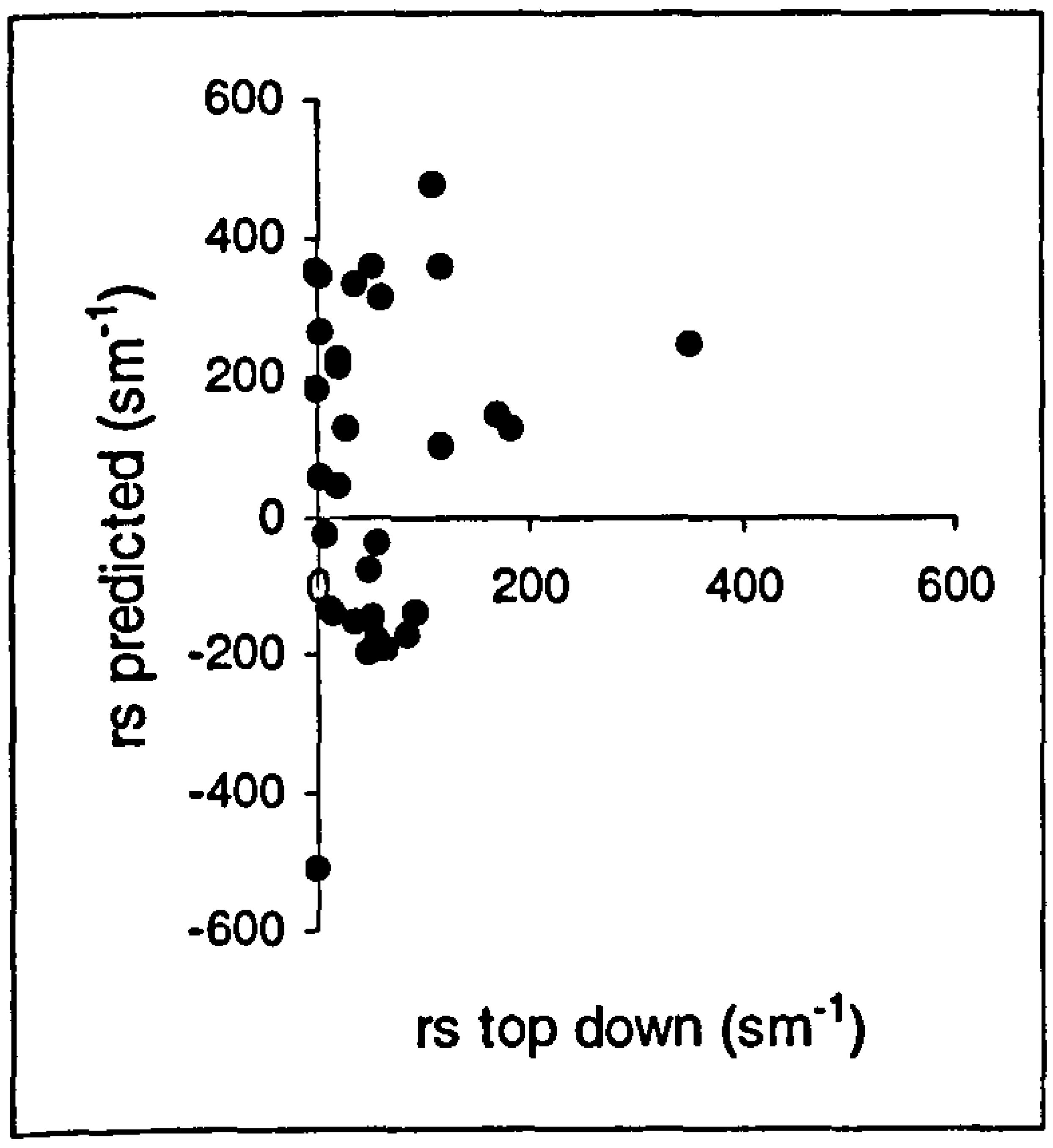

Figure F.07 – Results of the Jarvis model prediction.

It can be seen that the results are not good, often resulting in negative values when they should be positive, At first glance this result is surprising given the strength of the relationships in Figures F.02 – F.03 and F. 05 - F.06

F.3.3 Sensitivity analysis

The poor result can be explained by examining the sensitivity of the equation to parameters *e* and *f* (Figures F.08 and F.09).

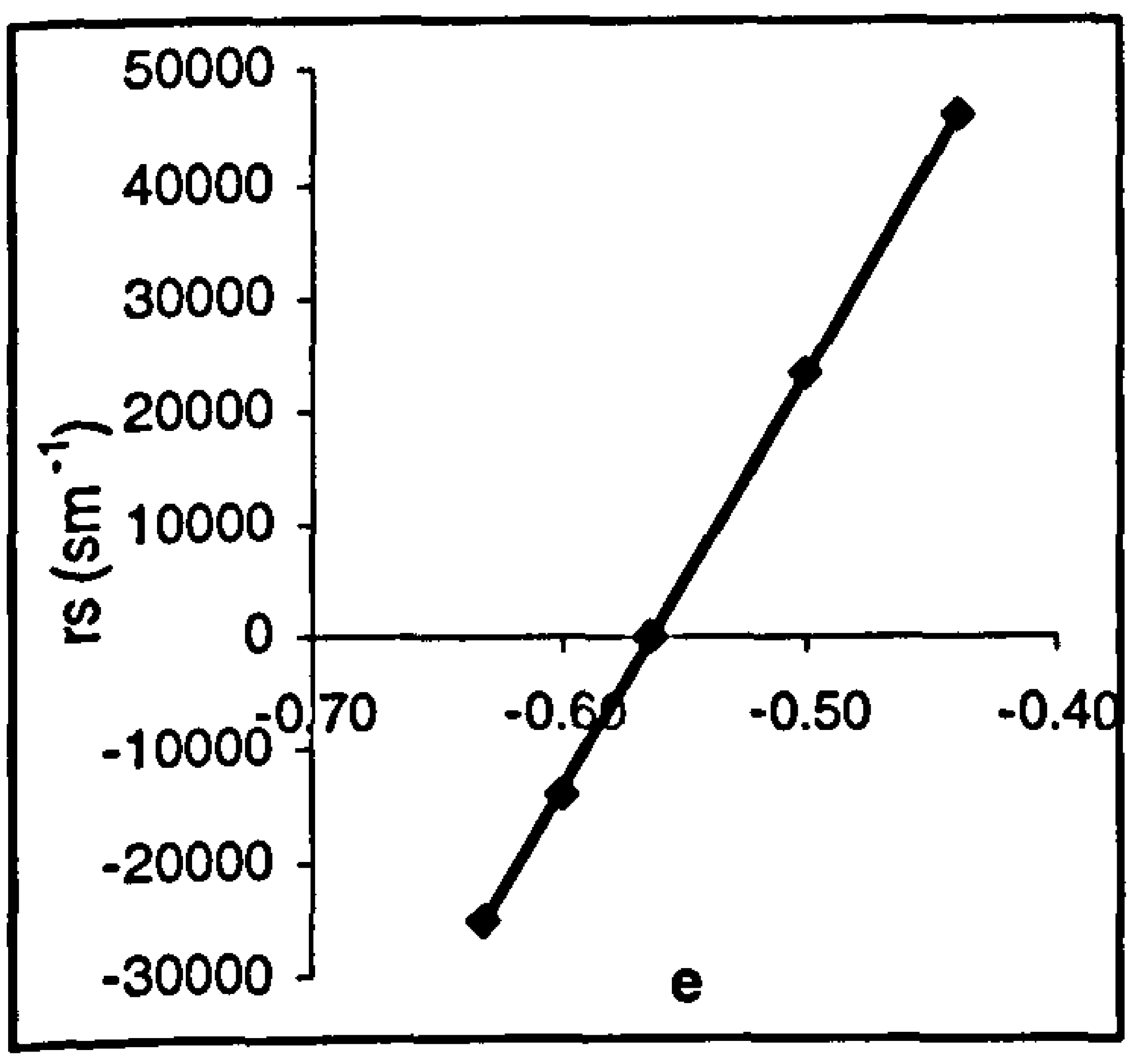

Figure F.08 – Sensitivity to parameter e

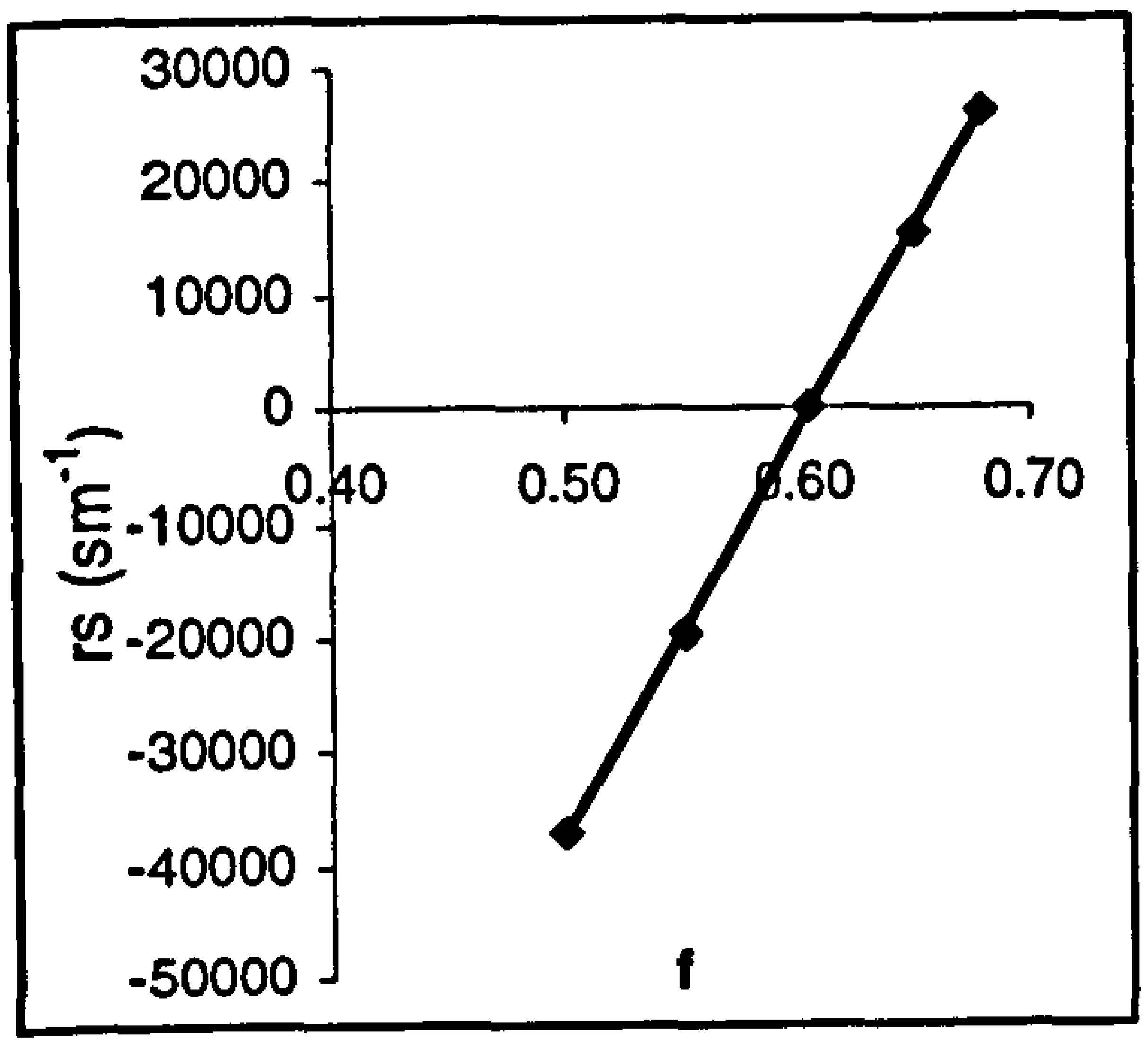

Figure F.09 – Sensitivity to parameter f

The equations are extremely sensitive to these parameters. A change of 0.001 in parameter *e* results in a change in estimation of stomatal conductance of 349.6 and for parameter *f* a change of 0.001 results in a change in estimation of stomatal conductance of 376.2. This means despite the goodness of the fit of Figures F.05 and F.06, even a tiny mis-estimation results in an overall huge error in the estimation of stomatal resistance.

F.4 DISCUSSION AND CONCLUSIONS

When using top down stomatal resistance data the functions of each parameter are effectively defined by the form of the Penman Monteith equation. Therefore those that were found in this study are the same as those used by Alves (2000) and other workers. However there are a large number of parameters that must be defined empirically and it was impossible to do this with any kind of success. The model became extremely sensitive to some parameters so that huge errors developed. In the end it proved impossible to create a working model of stomatal conductance with a consistent parameter set that could be used predictively. The problem of variable parameters was also found by Jarvis (1976) and Wever (2002).

Appendix G - Comparison of Meteorological Data from Stodmarsh and Manston

Daily averages of meteorological data collected from Manston Airport (51.349°N, 1.353°E, elevation 44 m) and Stodmarsh National Nature Reserve (51°19'N, 1°12'E, elevation <5 m) from 31/05/01 to 28/08/01 were compared in order to assess the validity of using Manston meteorological data to create a model of evapotranspiration at Stodmarsh, 10km away.

G.1 TEMPERATURE

Figure G.01 shows a comparison of temperature at Manston with temperature measurement (made at 2.48 m) at Stodmarsh.

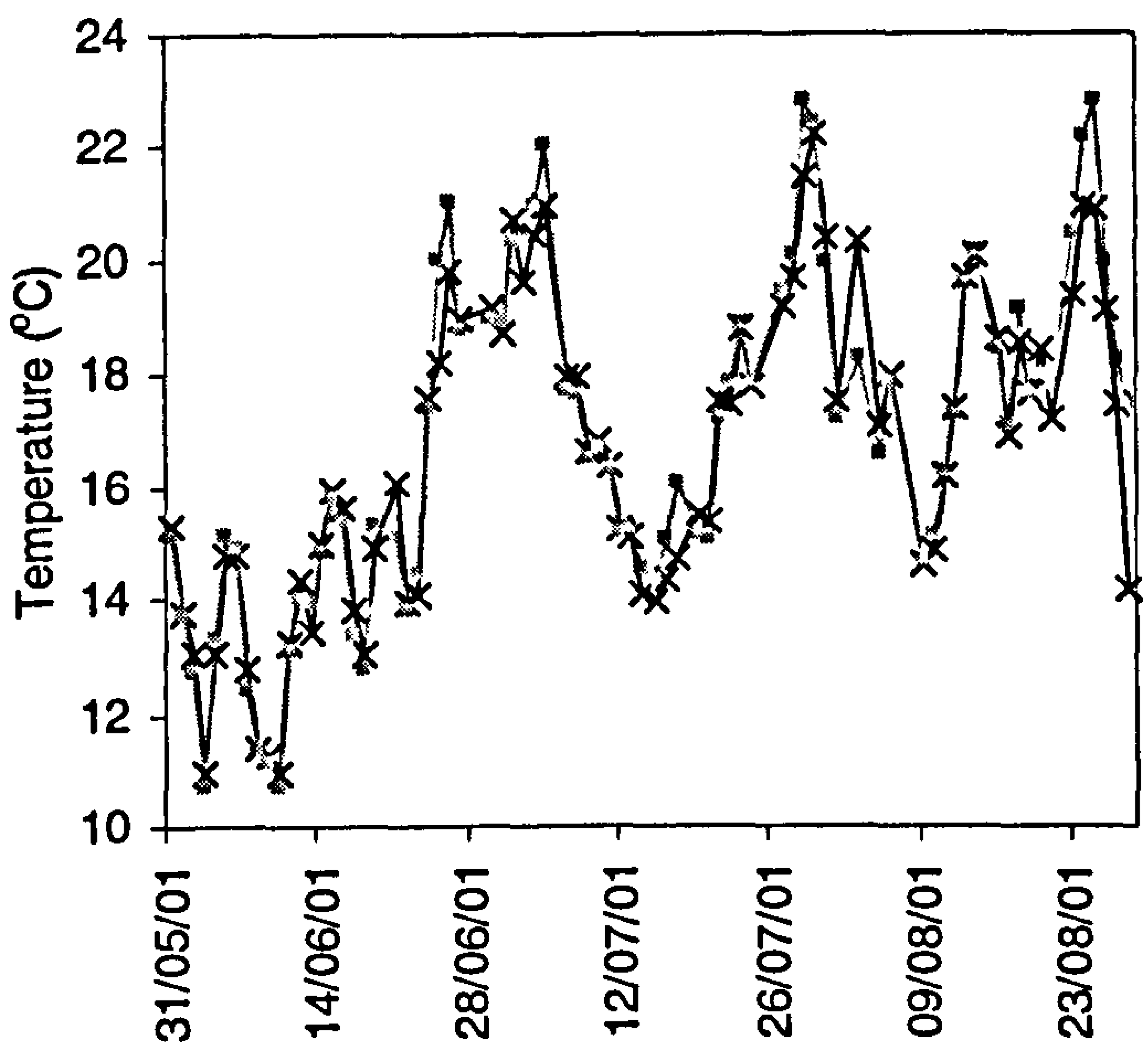

Figure G.01 – A comparison of daily mean temperature at Stodmarsh (black lines with crosses) and Manston (grey lines with squares).

With the exception of a few days, the two data sets are very similar.

G.2 WINDSPEED

As at all standard Meteorological Office stations, the windspeed at Manston is measured at a height of 10 m. This means that before comparison, it needs to be adjusted in order to be comparable to the height at which windspeed is measured at Stodmarsh (2.54 m). This adjustment is done following Allen *et al.* (1994b):

$$u_{2.54} = u_{10} \frac{\ln\left(\dfrac{2.54-d}{z_o}\right)}{\ln\left(\dfrac{10-d}{z_o}\right)} \qquad \text{(G.01)}$$

where $u_{2.54}$ is the mean windspeed measurement at Stodmarsh at 2.54 m and u_{10} is mean windspeed measurement at Manston at 10 m, z_o is roughness length governing momentum transfer (m) and d is the height of zero plane of displacement (m).

This equation also takes account of the differences in surface between the two sites – reeds at Stodmarsh and short grass at Manston. This is reflected in the differing values used for d and z_o. For Stodmarsh an average of the measured values is used (d = 1.108, z_o = 0.426 – see Chapter 4) and for Manston the standard values for short grass as given by Allen *et al.* (1994b) are used (d = 0.08, z_o = 0.015)

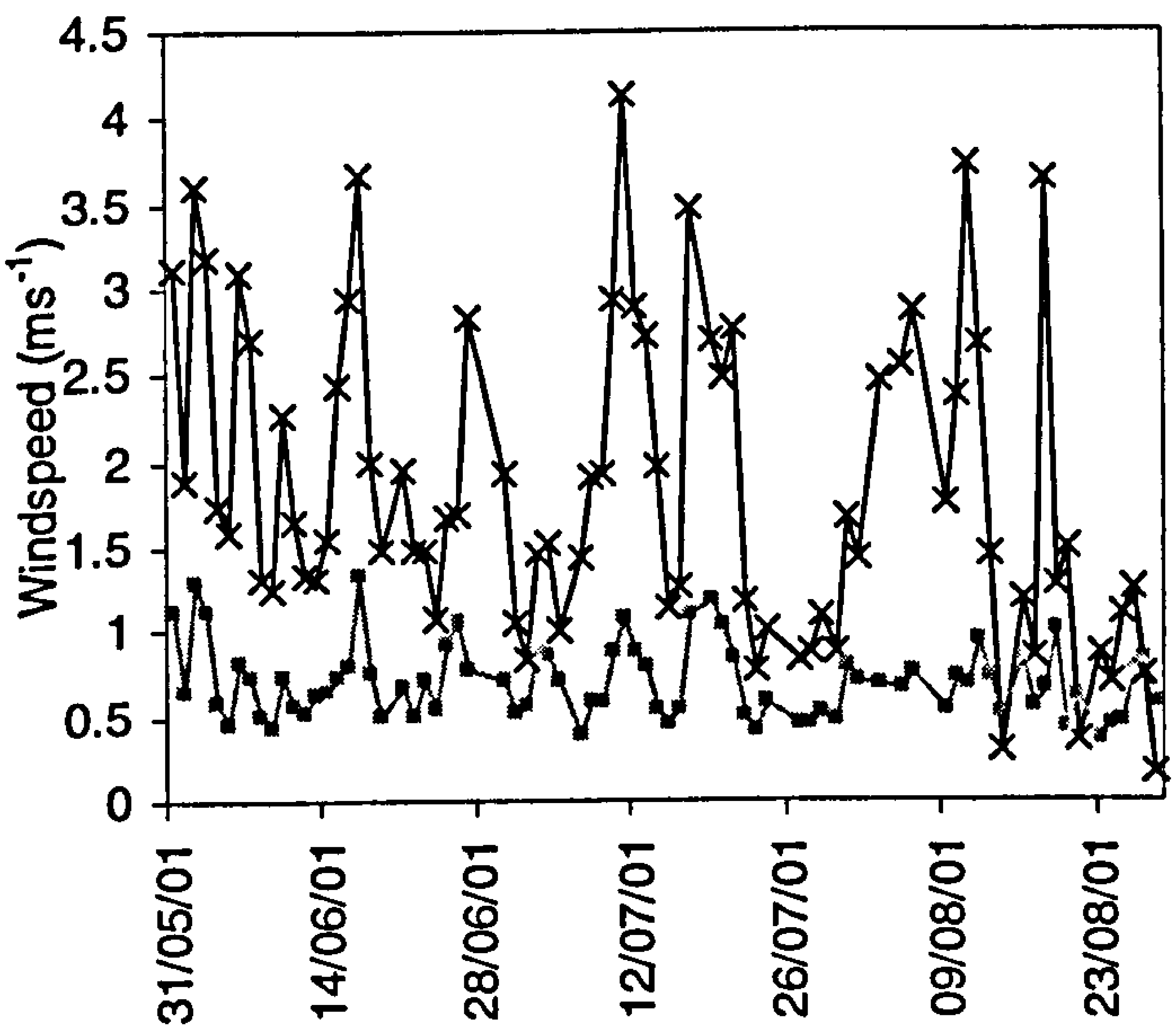

Figure G.02– Comparison of daily mean windspeed at Stodmarsh (black lines with crosses) and Manston (grey lines with squares).

Figure G.02 indicates that windspeeds at Manston were around half those at Stodmarsh. This seems unlikely in reality however as Manston is closer to the coast and at a higher elevation (44 m compared to less than 10 m). It appears the equation is suppressing windspeed too much as prior to the changes raw Manston data showed higher windspeeds than those at Stodmarsh. There is a linear relationship between the two measurements as seen in Figure G.03, albeit with lots of scatter.

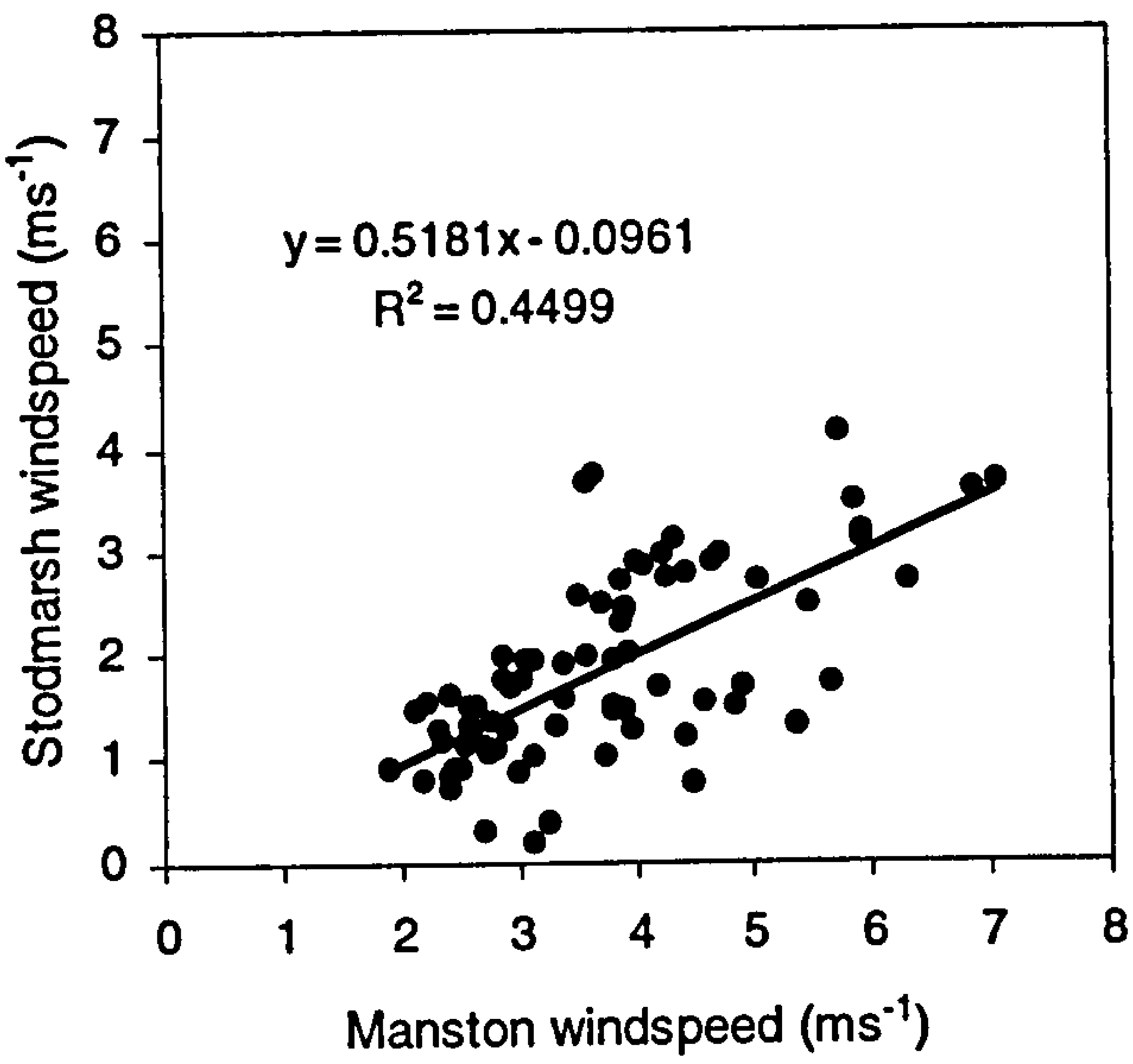

Figure G.03 – Relationship between windspeed measured at Stodmarsh and Manston

This relationship can then be used to bring Manston data to a magnitude that is similar to Stodmarsh. Though this is not perfect it is satisfactory.

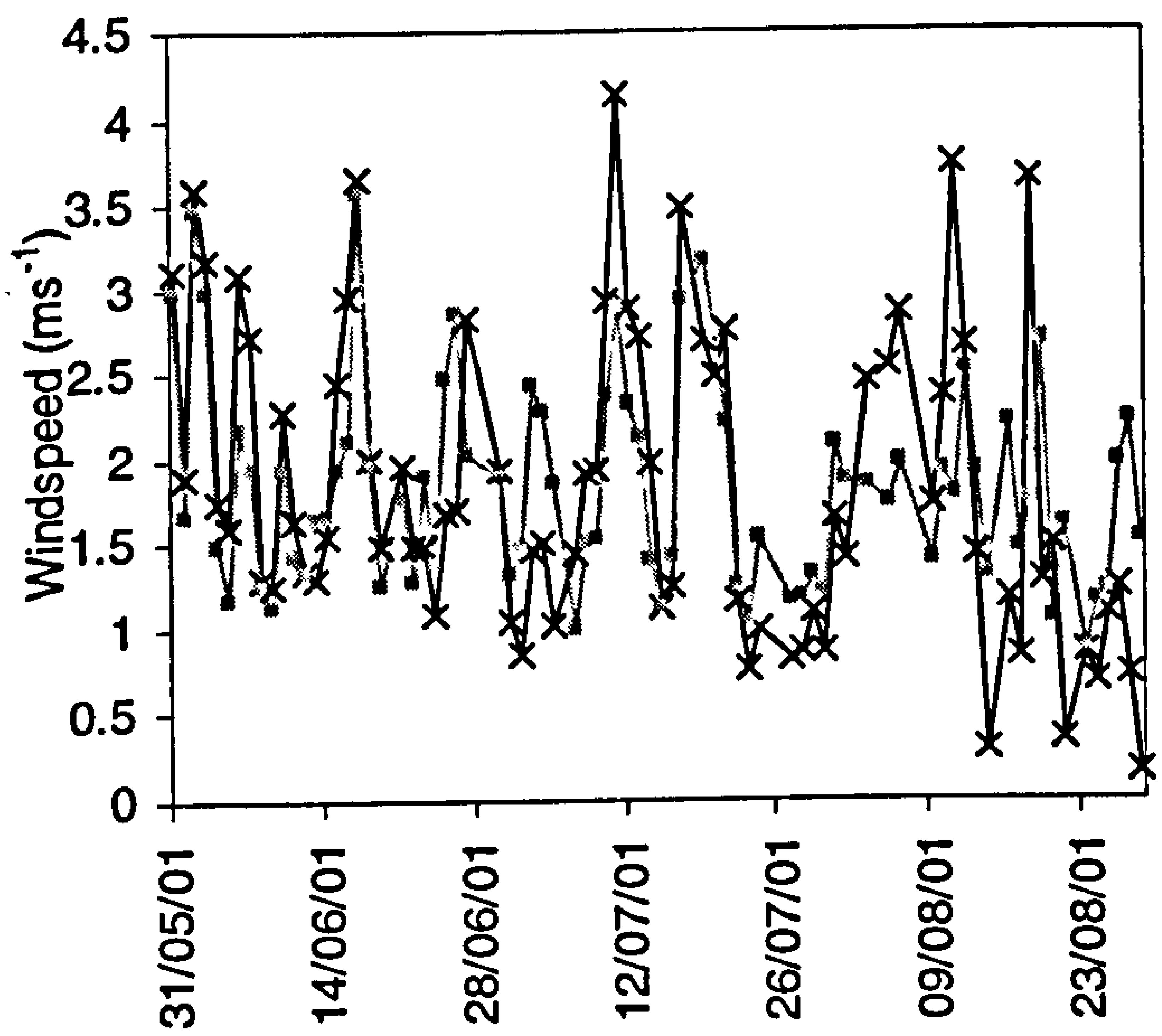

Figure G.04 – Windspeed at Stodmarsh (black lines with crosses) and adjusted windspeed at Manston (grey lines with squares) using linear regression.

G.3 NET RADIATION

Net radiation is more complex as it is not measured at Manston. From 1998 solar radiation (Rs) has been measured and for data before this, count of sunshine hours data must be used to calculate net radiation. When solar radiation is measured, net radiation is estimated from:

$$Rn = 0.77Rs - \left(a_c \frac{Rs}{Rso} + bc\right)\left(a_1 + b_1\sqrt{e_d}\right)\sigma \frac{\left(T_{\max}^4 + T_{\min}^4\right)}{2} \quad \text{(G.02)}$$

where Rs is incoming solar radiation (W m^{-2}), Rso is short wave radiation for a clear sky day (W m^{-2}), a_c and b_c are calibration parameters (1.35 and -0.35), a_l and b_l are correlation coefficients (0.34 and 0.14), T_{max} is maximum daily temperature (K), T_{min} is minimum daily temperature (K) and σ is the Stefan-Boltzmann constant (5.67 x 10^{-8} W m^{-2} K^{-1}) where:

$$Rso = \left(0.75 + 2\times10^{-5} z\right)Ra \quad \text{(G.03)}$$

where Ra is extra terrestrial radiation (W m^{-2})

$$Ra = \frac{24\times60}{\pi} G_{sc} d_r (\omega_s \sin\varphi \sin\delta + \cos\varphi \cos\delta \sin\omega_s) \quad \text{(G.04)}$$

where G_{sc} is the solar constant, Dr is the relative distance earth-sun, δ is the solar declination angle (rad), φ is latitude (rad) (0.893) and ω_s is sunset hour angle (rad).

This resultant calculated net radiation for Manston is compared with measured net radiation at Manston in Figure G.05 below.

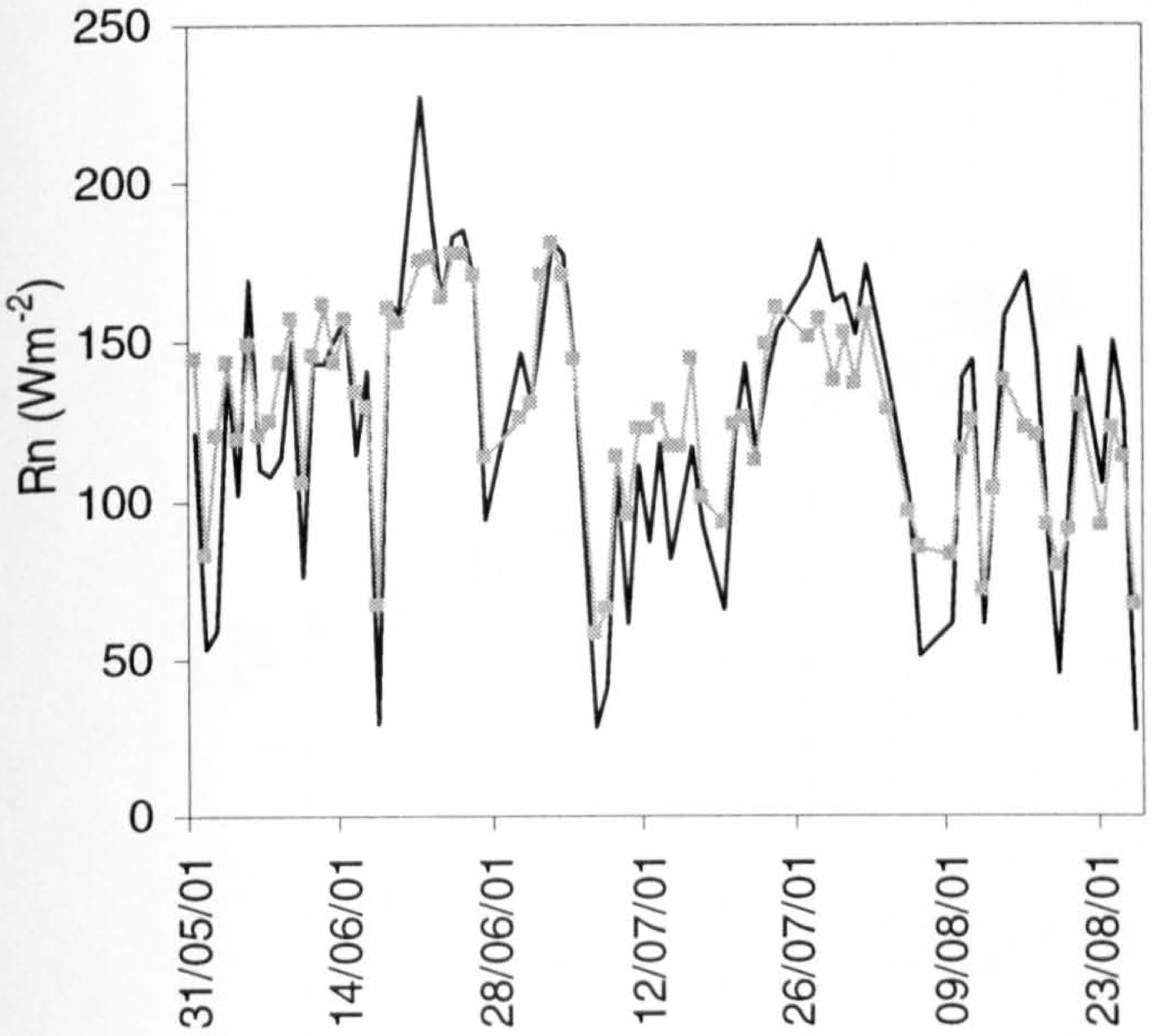

Figure G.05 – A comparison of net radiation from Stodmarsh (black lines with crosses) with net radiation, estimated from solar radiation from Manston (grey lines with squares).

When historical years' data from Manston are used in order to model evapotranspiration (Chapter 6) it was necessary to calculate net radiation from sunshine hour data using the equation:

$$Rn = 0.77\left(a_s + b_s \frac{n}{N}\right)Ra - 2.45\times10^{-9}\left(0.9\frac{n}{N} + 0.1\right)\left(a_1 + b_1\sqrt{e_d}\right)\left(T^4_{max} + T^4_{min}\right) \quad (G.05)$$

where a_s and b_s are angstrom constants, n is the number of bright sunshine hours per day and N is the total day length (hours).

Angstrom constants depend on site, pollution and time of year. They can be found from a regression of *n/N* against *Rs/Ra* as shown in Figure G.06. The intercept of the regression line is a_s and the slope is b_s. This results in constants of $a_s = 0.2677$ and $b_s = 0.4859$ which are close to the recommended average values of $a_s = 0.25$ and $b_s = 0.50$.

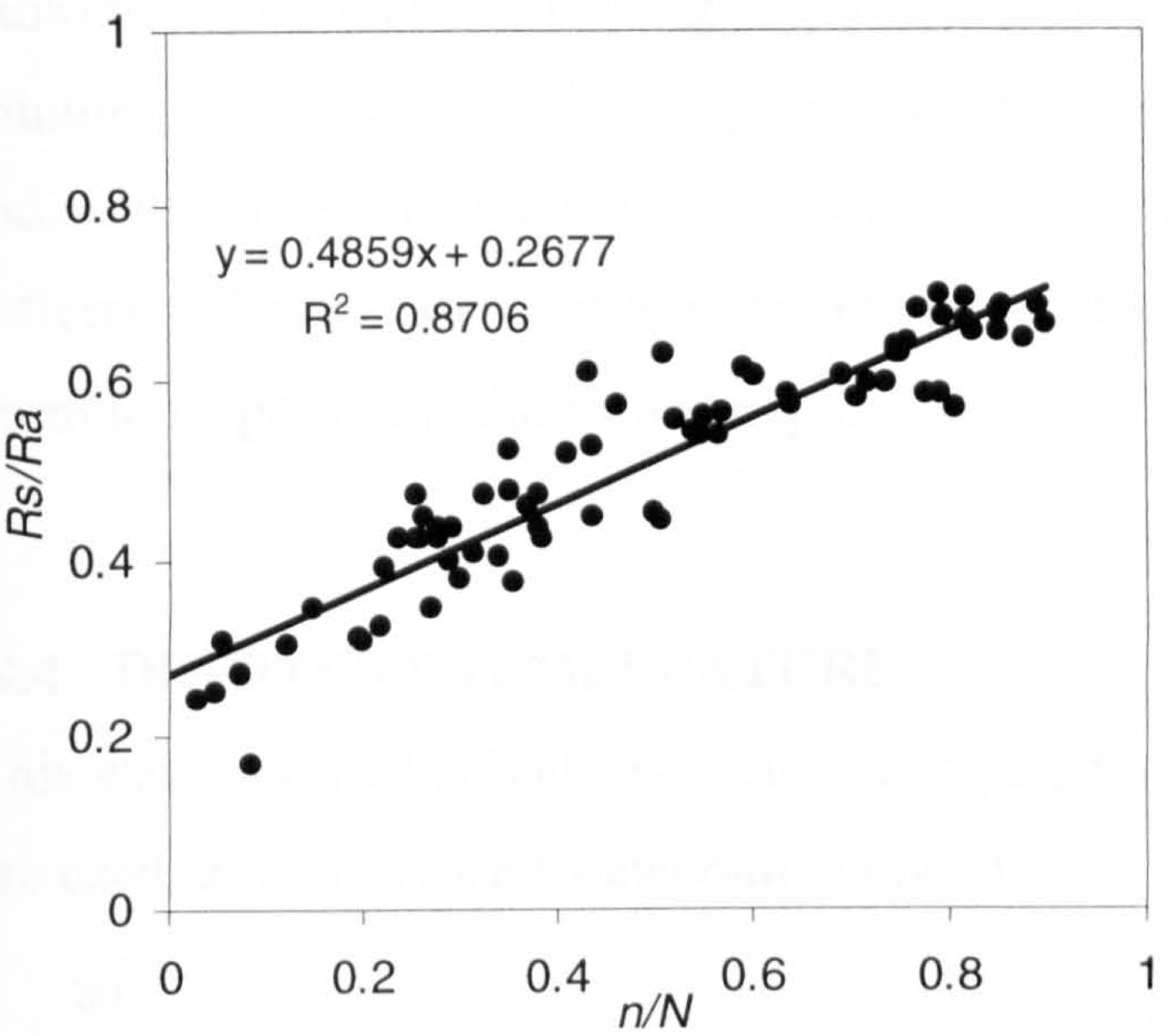

Figure G.06 – Regression of Rs/Ra against n/N in order to find Angstrom constants

These calculated Angstrom constants are used in the calculation of net radiation from sunshine hours. The results can be seen in Figure G.07.

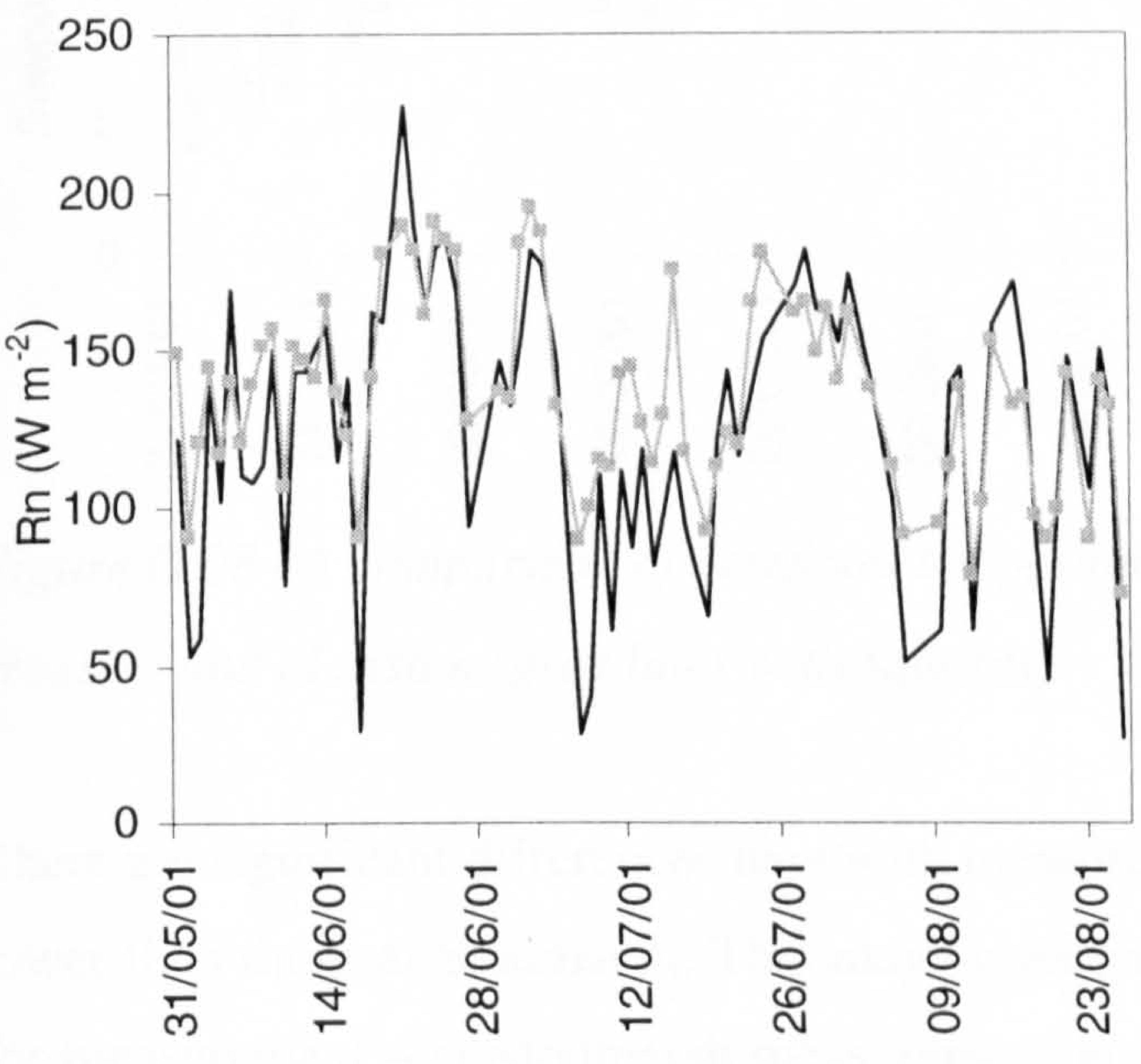

Figure G.07 – A comparison of net radiation from Stodmarsh (black lines with crosses) with net radiation, estimated from sunshine hours from Manston (grey lines with squares).

Although the Manston data does not always reach the same peaks as the Stodmarsh data, considering the sunshine hour data is a fairly crude measure compared to the net radiometer, which takes readings of both incoming and outgoing radiation every twenty minutes, the comparison is acceptable. There is potential for a difference in long wave radiation emission at the two sites due to the different surfaces which may have different albedos. However this should be revealed in a systematic error which does not seem to be present in the above graph.

G.4 DEWPOINT TEMPERATURE

This was measured at both sites and is compared in Figure G.08 (only daytime averages are used, as at Stodmarsh dewpoint temperature is only measured during the day).

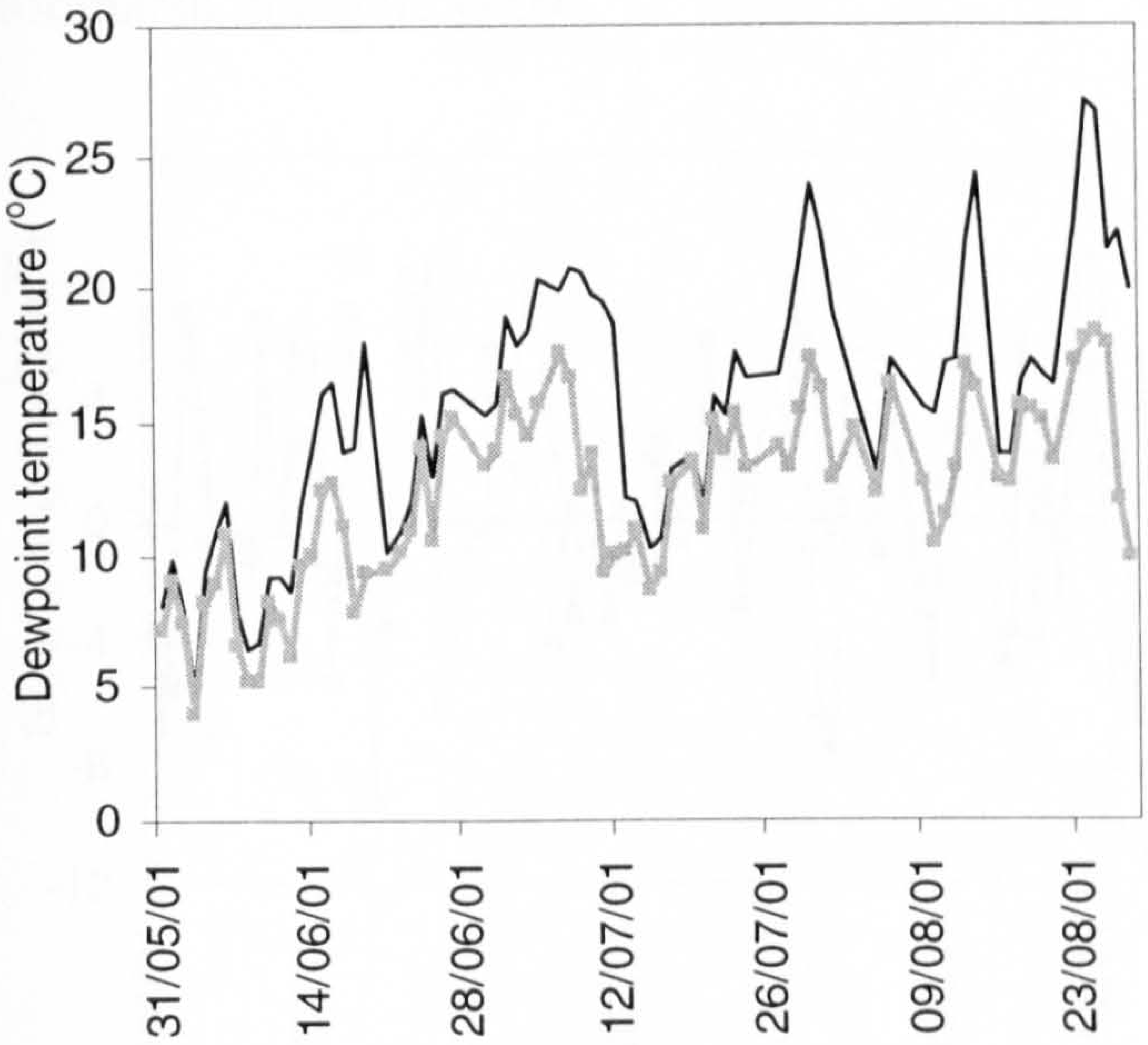

Figure G.08 –A comparison of dewpoint temperature at Stodmarsh (black lines with crosses) and Manston (grey lines with squares).

There are significant differences here with measurements at Manston being significantly lower than those at Stodmarsh. This may be partially due to the different methods used for measurement – at Stodmarsh measurement is made using a cooled mirror dewpoint hygrometer and at Manston it is made using a wet bulb thermometer and could be related to the different environments – humidity is likely to be higher above the transpiring wetland than above grass. Humidity measurements of all types are notoriously difficult to take accurately and this may be indicative of the fact that the measurements were not always correct.

G.5 GROUND HEAT FLUX

Ground heat flux is not measured at Manston but must be estimated from the equation:

$$G = \frac{(C_s z_s (T_i - T_{i-1})) * 100000}{86400} \qquad \text{(G.06)}$$

where C_s is soil heat capacity (2.983 MJ m^{-3}°C^{-1} - see Appendix A), z_s is effective soil depth (0.08 m),T is air temperature (°C) and T_{-1} is air temperature on previous day (°C)

The results of this are compared with mean ground heat flux measurements at Stodmarsh in Figure G.09

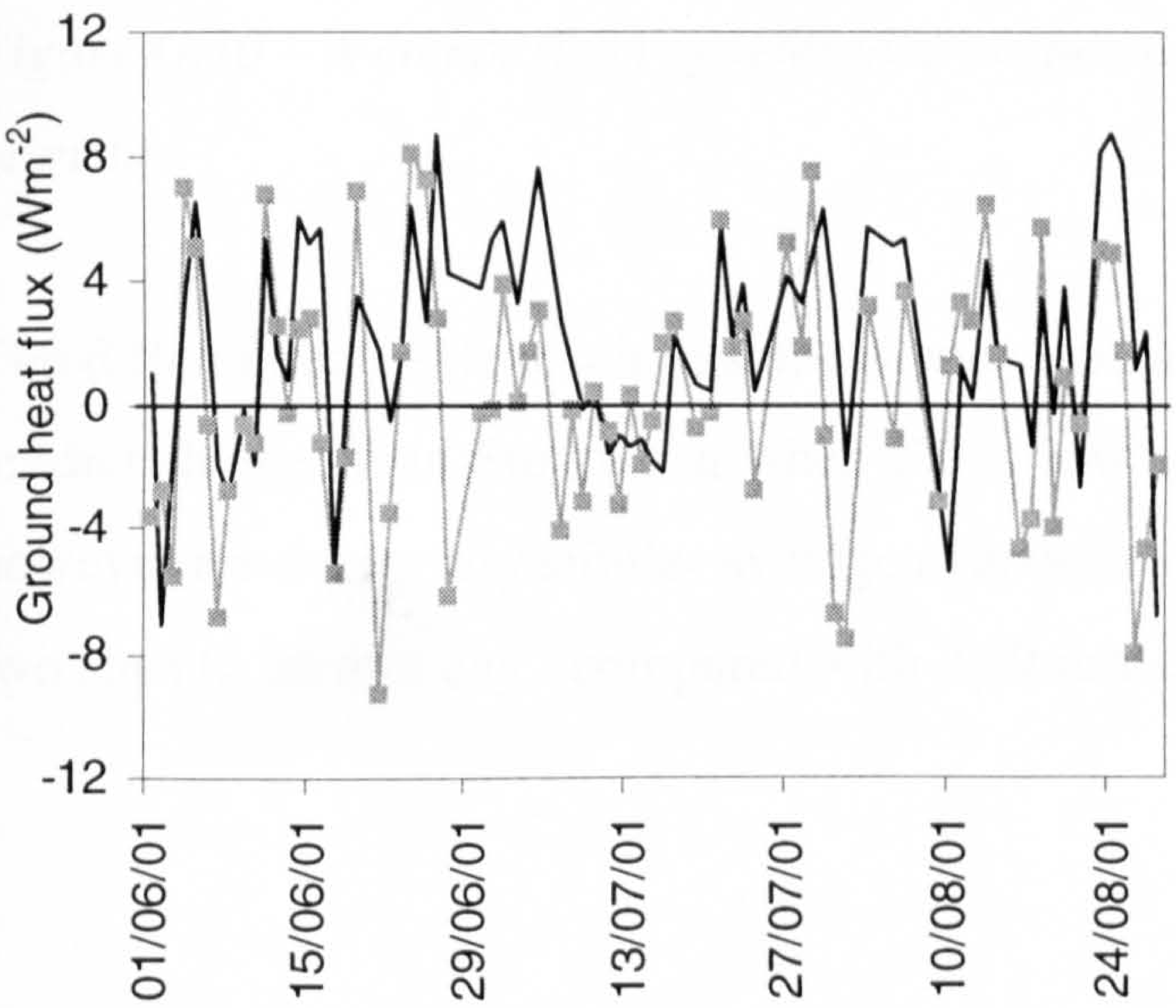

Figure G.09 – A comparison of ground heat flux estimations at Stodmarsh (black lines with crosses) and Manston (grey lines with squares).

Some parts of the data very fit well, others are less good. The fit should be adequate as on a daily basis, ground heat flux is generally less than 2% of net radiation.

G.6 OVERALL EVAPOTRANSPIRATION

The graph below compares reference evapotranspiration (Equation 3.14) calculated using data for Manston with reference evapotranspiration calculated using data from Stodmarsh for the same period.

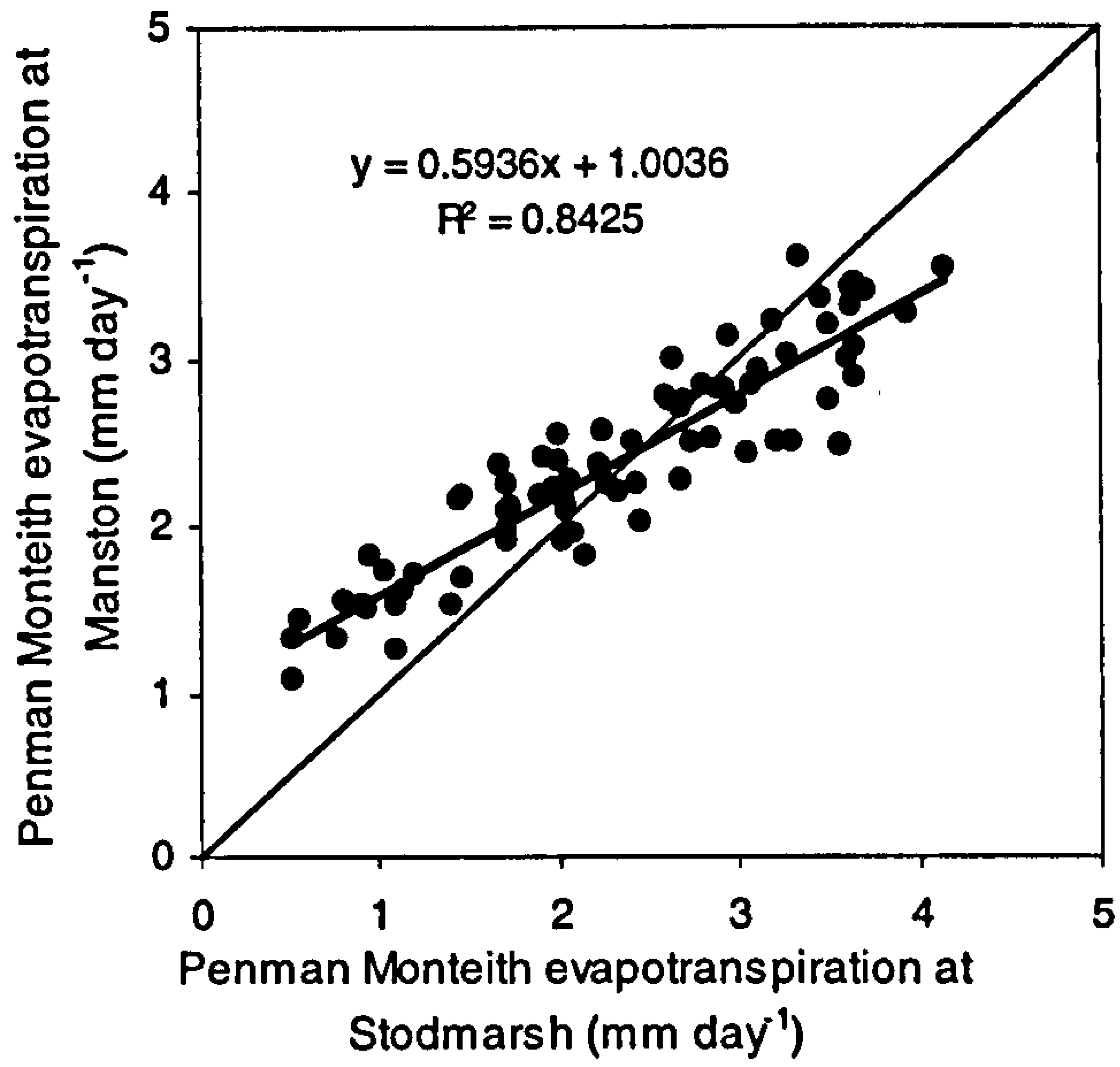

Figure G.10 – A comparison of reference evapotranspiration at Stodmarsh and Manston

The different data sets result in some scatter around the 1:1 line. It appears that Manston predicts higher than Stodmarsh when ET is low and vice versa when ET is high. This however results in very similar average evapotranspiration over the whole data set at the two sites (2.34 mm day^{-1} compared with 2.39 mm day^{-1}).

Appendix H – Calibration and Validation of Alternative Outflow Models

H.1 INTRODUCTION

Due to the poor validation of the IHACRES outflow model and its over-prediction within the water balance some alternative modelling approaches were tried in order to attempt to improve upon it.

H.2 MODEL CALIBRATION AND VALIDATION

H.2.1 WaSim / Model Maker model

The WaSim / Model Maker model is a two step model and follows a similar approach to IHACRES in that the initial step determines effective rainfall and the second how this effective rainfall results in streamflow leaving the reserve. Effective rainfall is determined using WaSim, a model principally aimed at simulating sub-surface drainage. It uses data inputs of daily rainfall and reference evapotranspiration. It allows soil characteristics for the catchment to be set (water retention, hydraulic conductivity, curve number and the depth of the soil profile) as well as crop factors including planting and harvest dates. These factors were initially set from knowledge of the catchment and refined through calibration. There are obvious difficulties with choosing single figures for these factors in a diverse catchment, especially for the outflow when there is likely to be big variations between the arable land upstream and the area of the reserve. Therefore the parameters are not necessarily representative of reality.

WaSim creates an output of thoughflow through the soil and of surface runoff as a result of rainfall. These flows are then used as an input into a model created in Model Maker. A schematic of the model can be seen in Figure H.01.

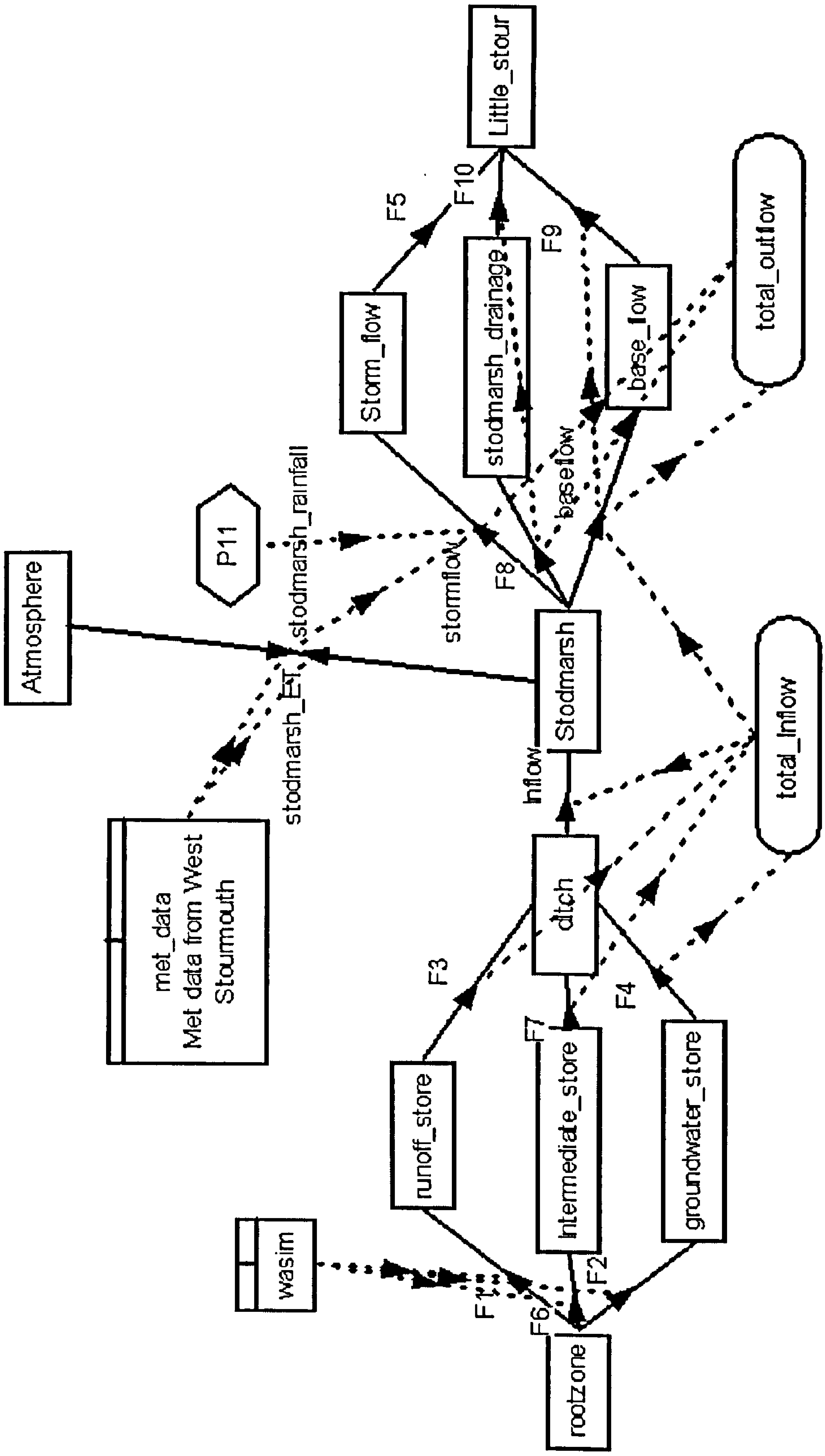

Figure H.01 – Schematic diagram of the ModelMaker model

(Words in italics refer to labels or parameters in Figure H.01). The *root-zone* represents the ground on which the effective rainfall falls upstream of the reserve. WaSim separates effective rainfall into runoff, which flows into the *runoff store*, and drain flow which is divided between the *intermediate store* and *groundwater store*, the proportions varying according to the parameter *gw*. All three flows are multiplied by separate empirical parameters (k_{runoff}, $k_{intermediate}$ and $k_{groundwater}$) and the sum of the flows is assumed to represent *total inflow* to the site. The subsequent part of the model represents flow through the reserve. This is influenced by rainfall and evapotranspiration (*met data from West Stourmouth*) directly onto and from the site. Again the flow is divided into three components. If rainfall is greater than *P11* rainfall goes into the *Storm flow* component, *Stodmarsh drainage* has a constant value and *Base flow* is total inflow minus k_4. The *total outflow* is the sum of the three flows. All parameters are determined empirically during calibration in order to get the best match with measured streamflow data.

H.2.1.1 WaSim / IHACRES model - Calibration and validation

The calibrated constants for the model are shown in Table H.01.

Table H.01 – Values used for each constant parameter of the WaSim / ModelMaker model

Constant	value
gw	0.2
k_{runoff}	0.3
$k_{intermediate}$	1
$k_{groundwater}$	0.004
P11	20 mm
Stodmarsh drainage	3500 $m^3 d^{-1}$
k_4	900 $m^3 d^{-1}$

The calibration period used was much shorter than for the IHACRES calibration. The comparison between modelled and measured outflow is seen in Figure H.02 below.

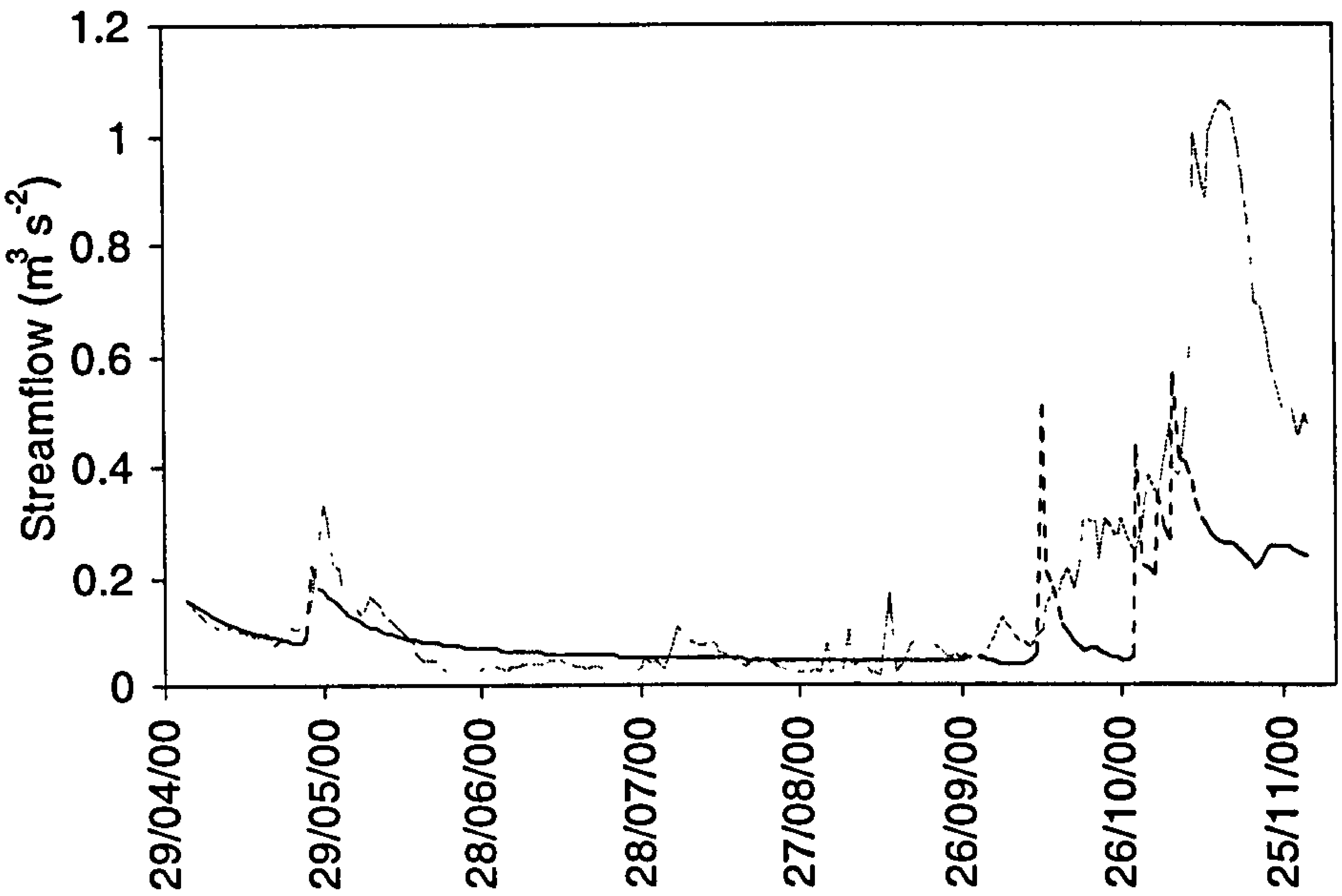

Figure H.02 – Calibration of modelled outflow using the WaSim / Modelmaker model against observed streamflow. The grey solid lines show observed data and the black dashed lines modelled data.

Figure H.03 below shows the validation of the WaSim / Model Maker model over the whole measurement period.

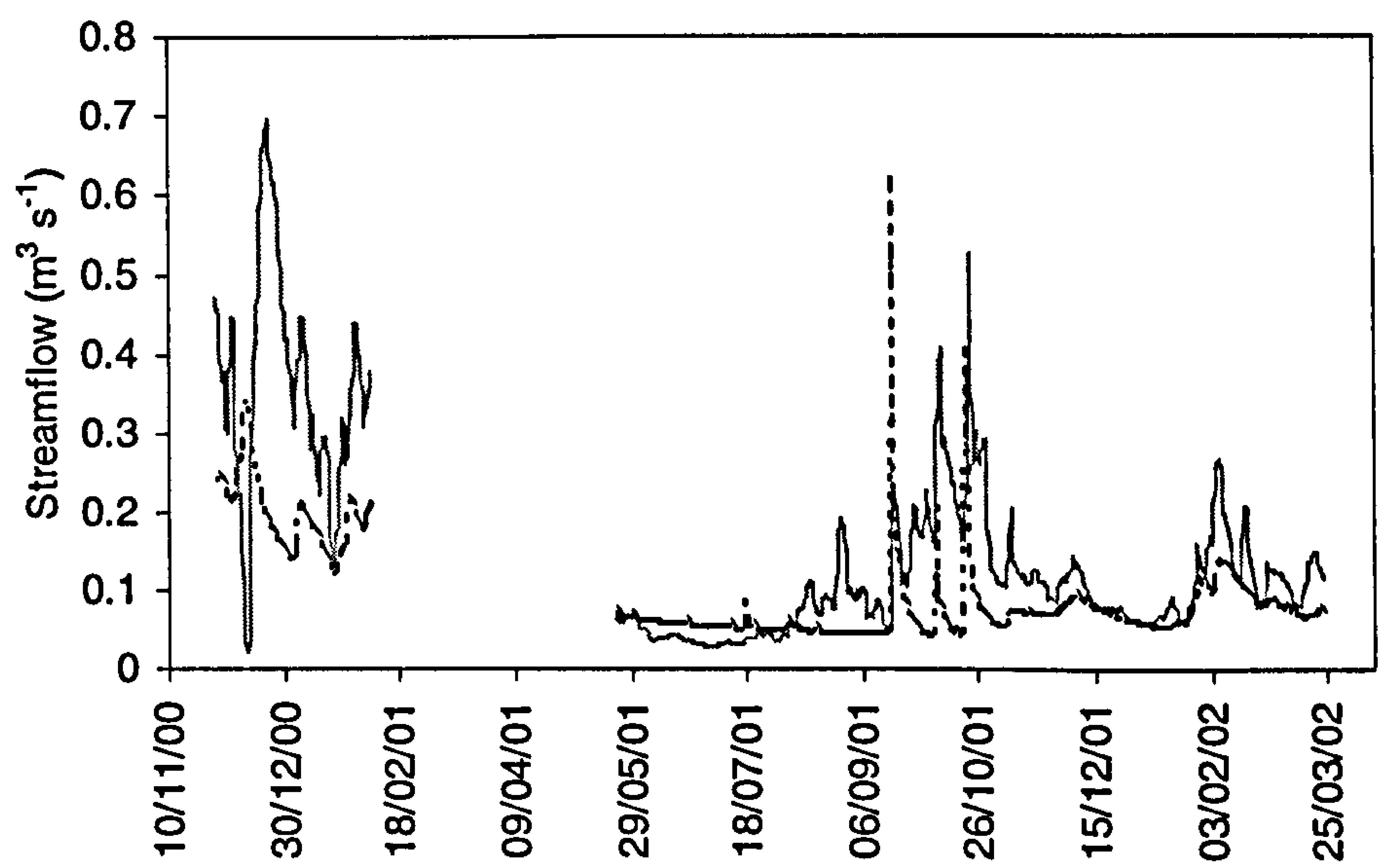

Figure H.03 – Calibration of modelled outflow using the WaSim / Modelmaker model against observed streamflow. The grey solid lines show observed data and the black dashed lines modelled data.

The same goodness of fit statistics were used as described to validate the IHACRES model in Chapter 6 (6.4.2.1), with the exception than R is not included as the calibration period was too short to create monthly data. The values can be seen in Table H.02 below.

Table H.02 - Goodness of fit statistics for the calibration and validation of the WaSim / ModelMaker model.

Test	Calibration	Validation
D_c	0.403	0.297
E	0.326	0.179
X^2 of sign test	0.46	57.81

Both in calibration and validation the fit of the WaSim / Model Maker model to the observed data is not as good as that of the IHACRES model. The fit is particularly poor in November when flows are very high.

H.2.1 Outflow regression model

It was noted that there was a relationship between observed inflow and observed outflow. Whether this could be exploited to create an outflow model was investigated.

H.2.1.1 Calibration

Simple linear regression was carried out between the inflow and outflow data (Figure H.04). Some extreme points which were very influential in the regression equation were removed.

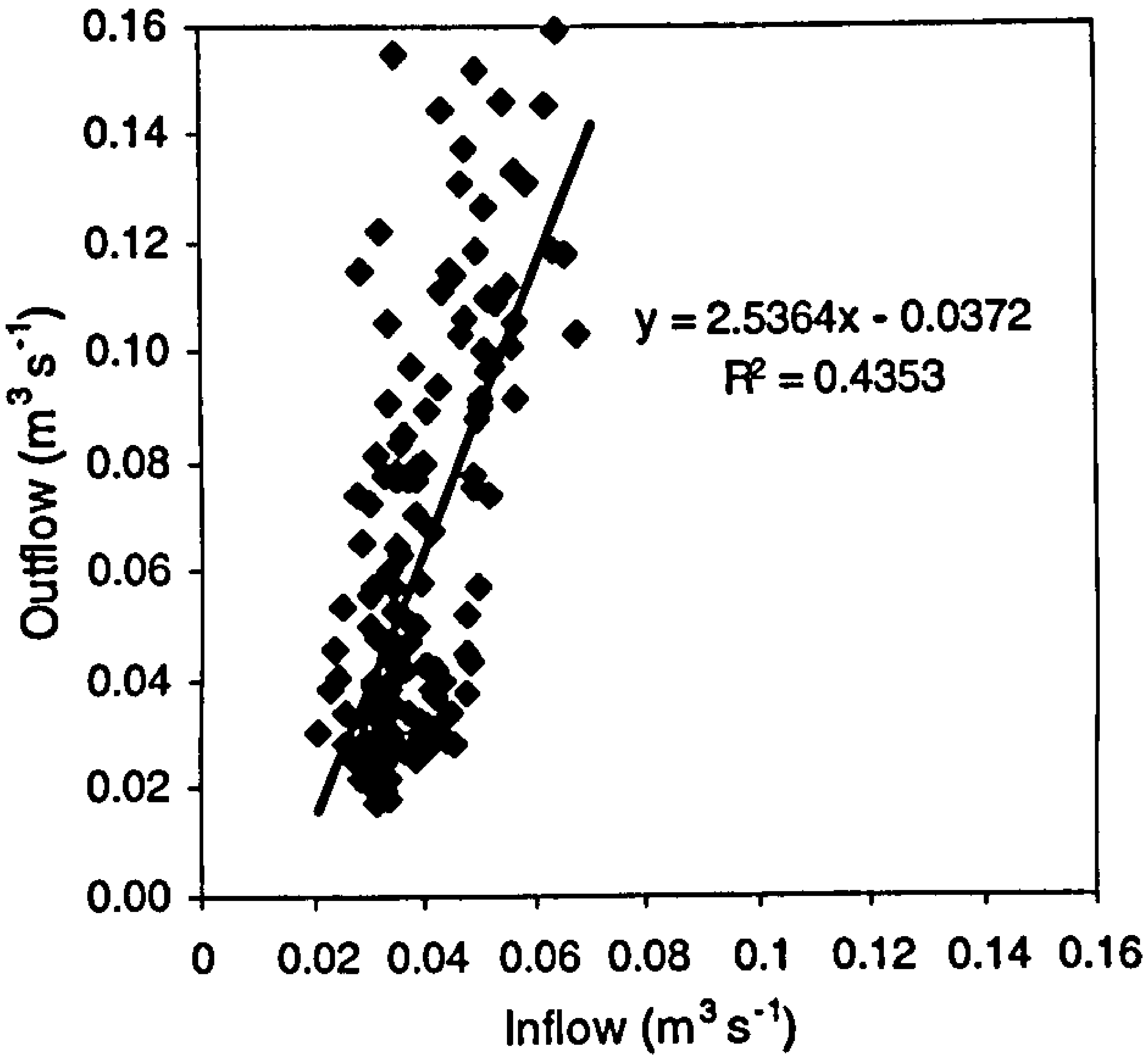

Figure H.04 – The relationship between observed inflow and outflow data.

The regression equation was used to model outflow over the part of the period for which there is observed data. The plot of the calibration can be seen in Figure H.05.

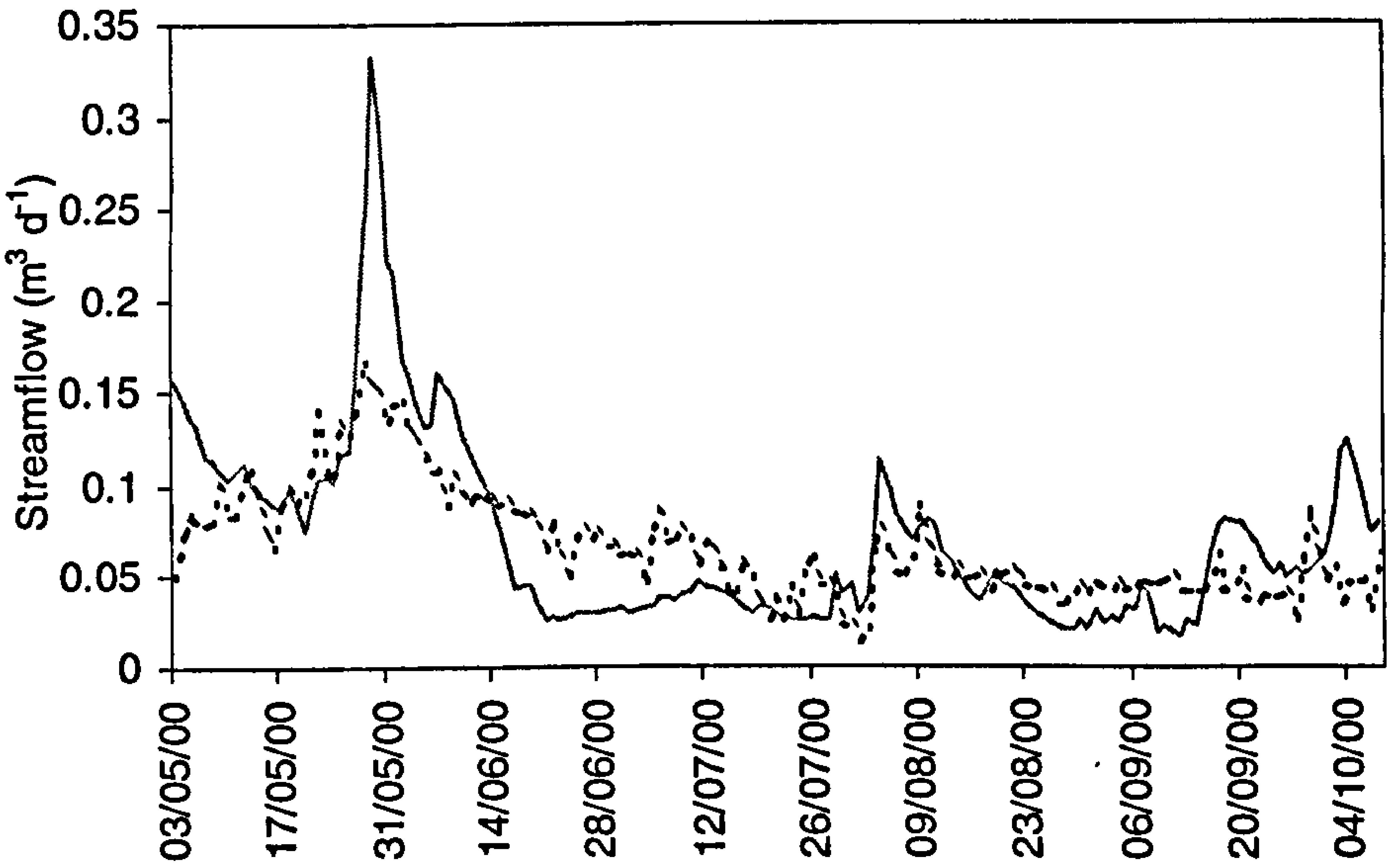

Figure H.05 – Calibration of modelled outflow using the regression model against observed streamflow. The grey solid lines show observed data and the black dashed lines modelled data.

The regression model was validated against an independent data set.

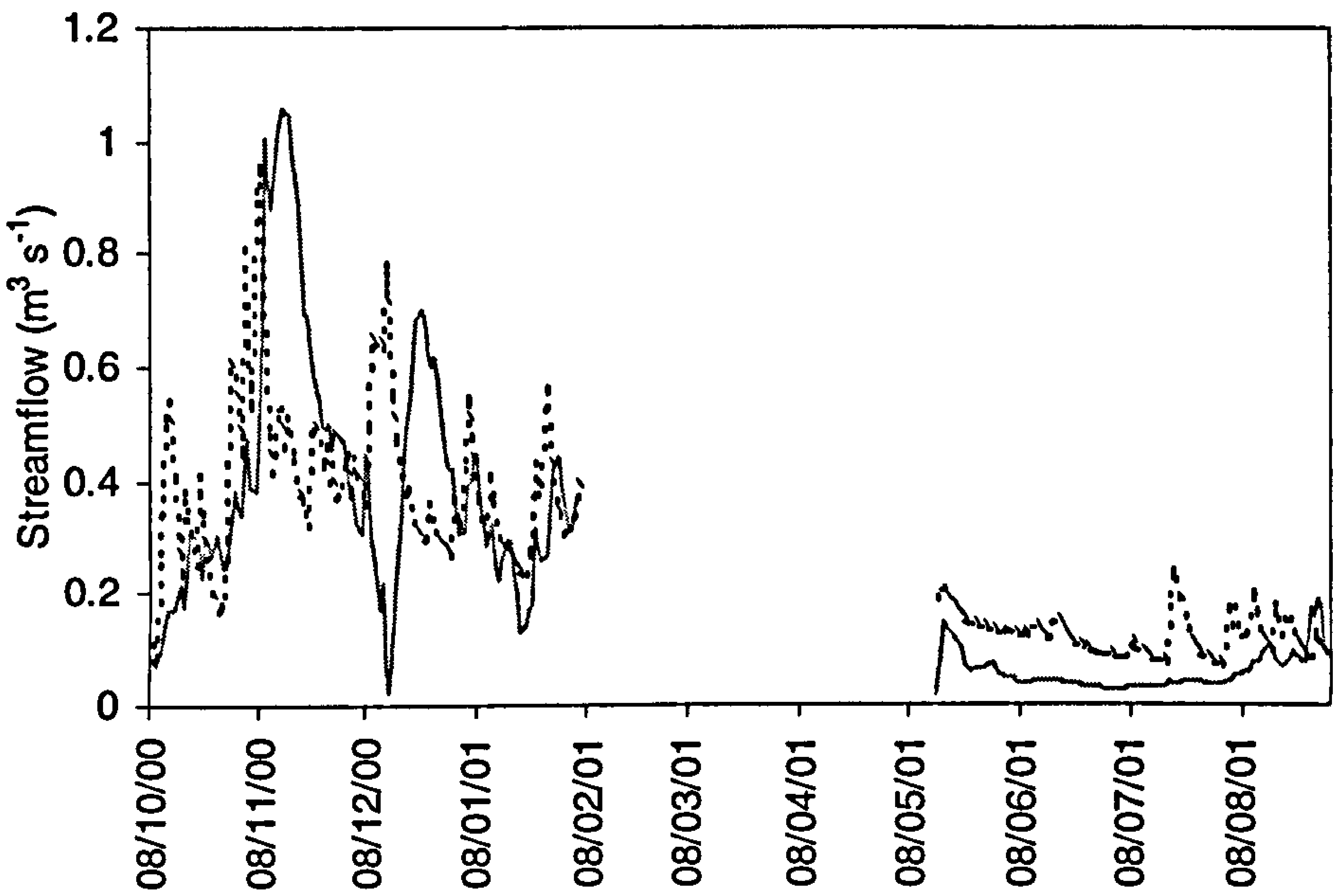

Figure H.06 – Validation of modelled outflow using the regression model against observed streamflow. The grey solid lines show observed data and the black dashed lines modelled data.

Table H.03 - Goodness of fit statistics for calibration and validation of the regression model.

Test	Calibration	Validation
D_c	0.814	0.323
E	0.461	-2.42
X^2 of sign test	2.29	43.84

H.3 CONCLUSIONS

Both models have calibration and validation statistics that are worse than those of the IHACRES model. No improvement is made. The WaSim/IHACRES model is the better of the two and is used in an experimental water balance model in Appendix I.

Appendix I – Water Requirement Predictions Using the Water Balance Model

I.1 INTRODUCTION

In Chapter 5 a simple water balance model was established as being appropriate to model hydrology on Stodmarsh National Nature Reserve. This model allows the change in storage to be calculated as a residual and can be used to predict actual water levels on the reserve - a very useful property. An attempt was made to use this model to predict water levels on site and to subsequently determine whether additional water may be required under different climatic regimes. This involved creating models of the water balance components. However due to the shortness and non-average conditions of the calibration period of the IHACRES outflow model, using this model within the water balance did not give sensible results in the water balance – over-predictions in the IHACRES model meant that the water balance appeared to predict large deficits of water which were unrelated to the amount of rainfall that had fallen. The water balance model was re-run using the WaSim/IHACRES model. The uncertainties in this model, indicated by its poor validation statistics, result in unacceptable uncertainties in the resulting water balance model and the predictions of surplus and deficits of water on the site. For this reason the following analysis was not included in the main body of the thesis but is included below as an example of the analysis that could be carried out were an improved outflow model, combined with other improvements in the understanding of site hydrology to become available.

I.2 METHODOLOGY

The water balance model used was:

$\Delta V/\Delta t = (P_n + S_I + R_v) - (ET + S_o)$ (after Mitsch and Gosselink 2000) (I.01)

where $\Delta V/\Delta t$ is change in volume of water storage in a wetland per unit time, P_n is precipitation, ET is evapotranspiration, S_I is surface inflow , S_o is surface outflow and R_v is river input.

Inflow was modelled using the IHACRES model, outflow with the WaSim/Modelmaker model, rainfall was from historical data and evapotranspiration was calculated from the

Penman Monteith equation using historical meteorological data from Manston Airport and crop coefficients from the Walton Lake site of Fermor *et al.* (2001). In the measured water balance it was found that input from the river was important in the water balance during winter months and this has been confirmed by the site managers own observations. This was accounted for by doubling stream inflow between November and March. The water balance was created from 1974-2001, using 1973 as a "warm up" period.

The water requirements of the site specify that the water should be at a depth of 200-300 mm in March. However if abstraction is to occur it must take place around the end of December. According to the model, the mean change in storage between January and March is an increase of 87 mm. Therefore the desired value for the beginning of January was set at 250 mm - 87 mm = 163 mm (395927 m^3 of water over the whole site). The difference between this figure and the storage level on 31st December of the same year was used to calculate the surplus or deficit of water that year. If there is a deficit, this is the amount that would need to be abstracted in order to regain target water levels.

I.3 RESULTS

The annual surplus or deficit was related to the annual rainfall (Figure I.01)

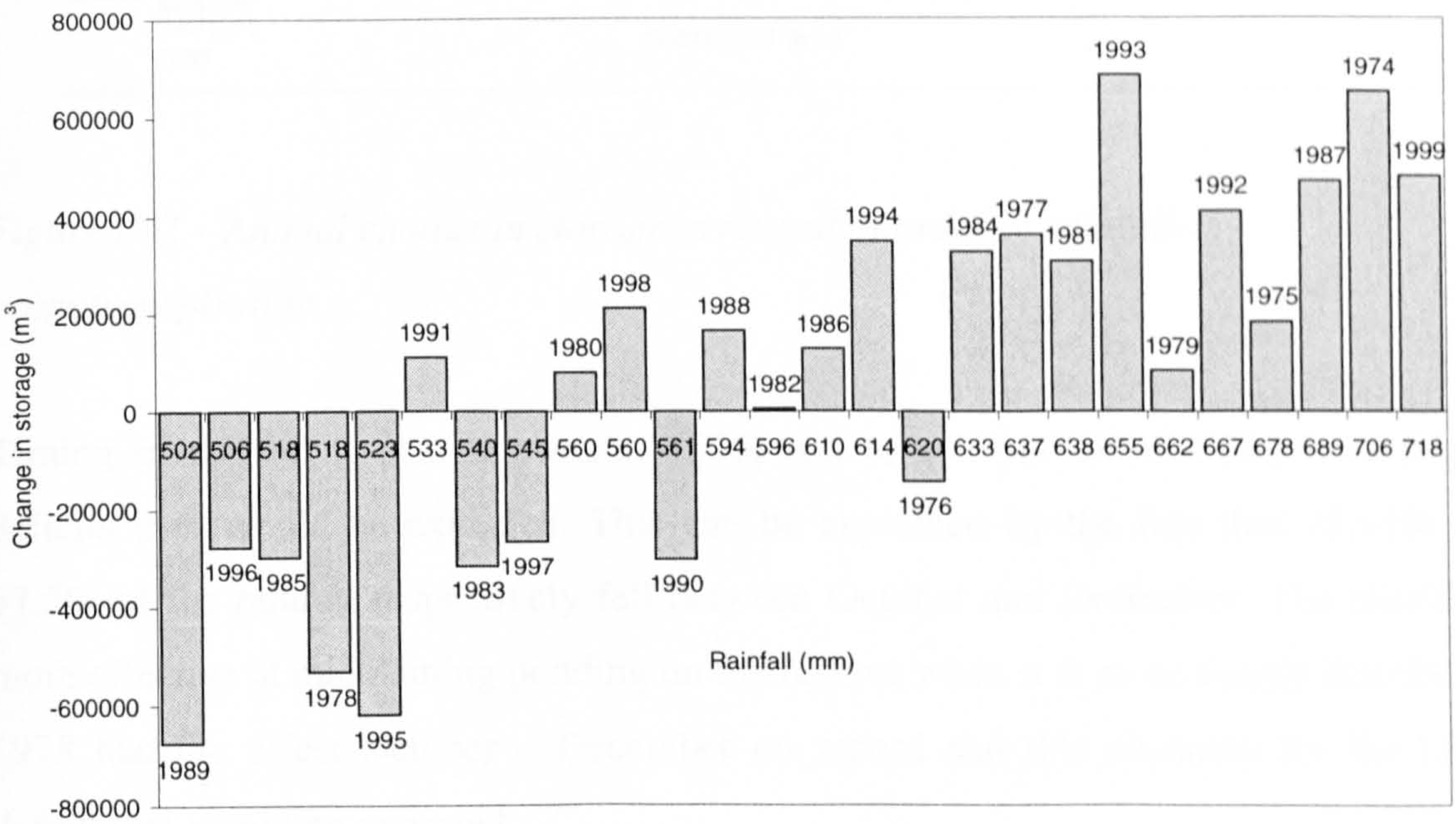

Figure I.01 – Deficit and surplus of annual water stored arranged in order of ascending total annual rainfall

An annual rainfall total of 523 mm or less always resulted in a deficit of water and greater than 633 mm always resulted in a surplus. In between these extremes there is some variation indicating that total annual rainfall is not the only factor determining the change in storage. Evapotranspiration rates are also a factor. For example 1990 has a bigger deficit than might be expected from the annual rainfall total and this is explained by high evapotranspiration and low summer and spring rainfall. An improved relationship can be created using rainfall minus evapotranspiration as shown in Figure I.02.

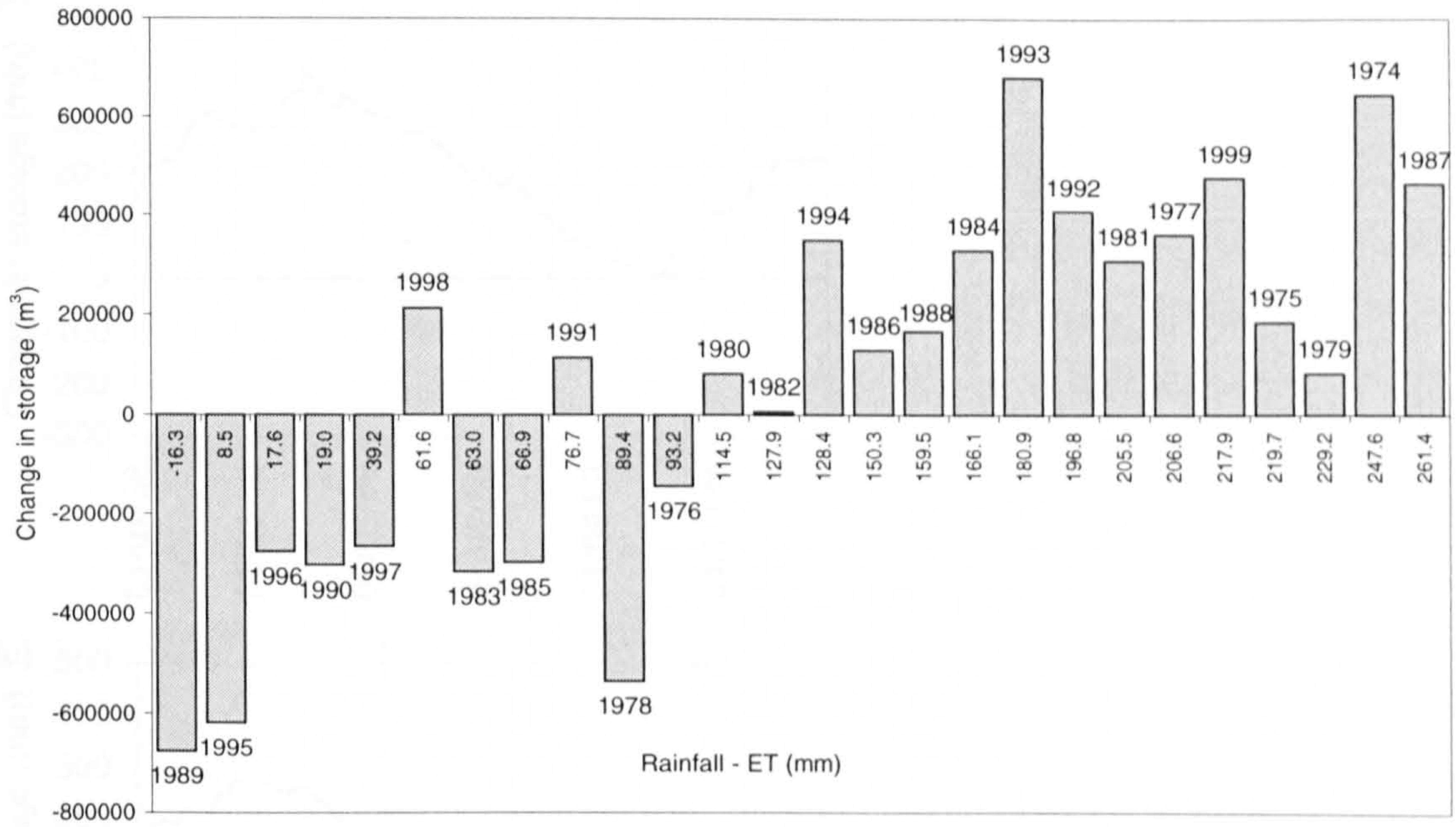

Figure I.02 – Annual change in storage arranged in order of rainfall-evapotranspiration.

Timing of rainfall is also important. 1976, and to a lesser extent 1982 have bigger deficits than would be expected. This can be explained by the fact that 75.51% and 57.5% of the rainfall respectively fell between October and December. The rainfall is more effective at maintaining ponding on the reserve when it is more evenly distributed. 1978 had the driest October – December on record and this accounts for the larger deficit that would be expected.

The graphs below show some examples of the annual water balances created.

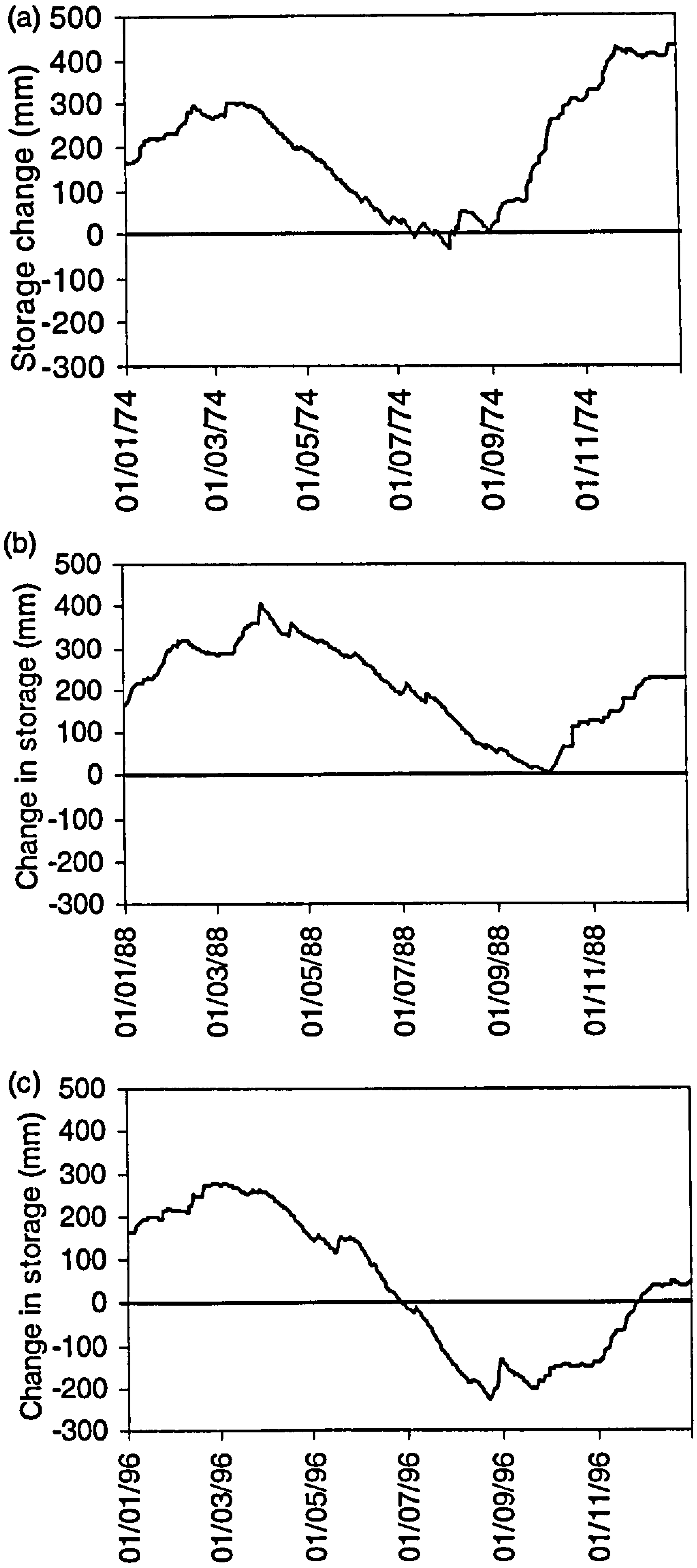

Figure I.03 – Storage on Stodmarsh National Nature Reserve in years with meteorological conditions equivalent to (a) 1974 – a wet year, (b) 1988 – an average year and (c) 1996 – a dry year.

The model can be tested to some extent by comparing the modelled data for 2000 with the measured storage. It can be seen in Figure I.04 that there is a reasonable agreement (the initial value is slightly different to account for the actual conditions in January 2000).

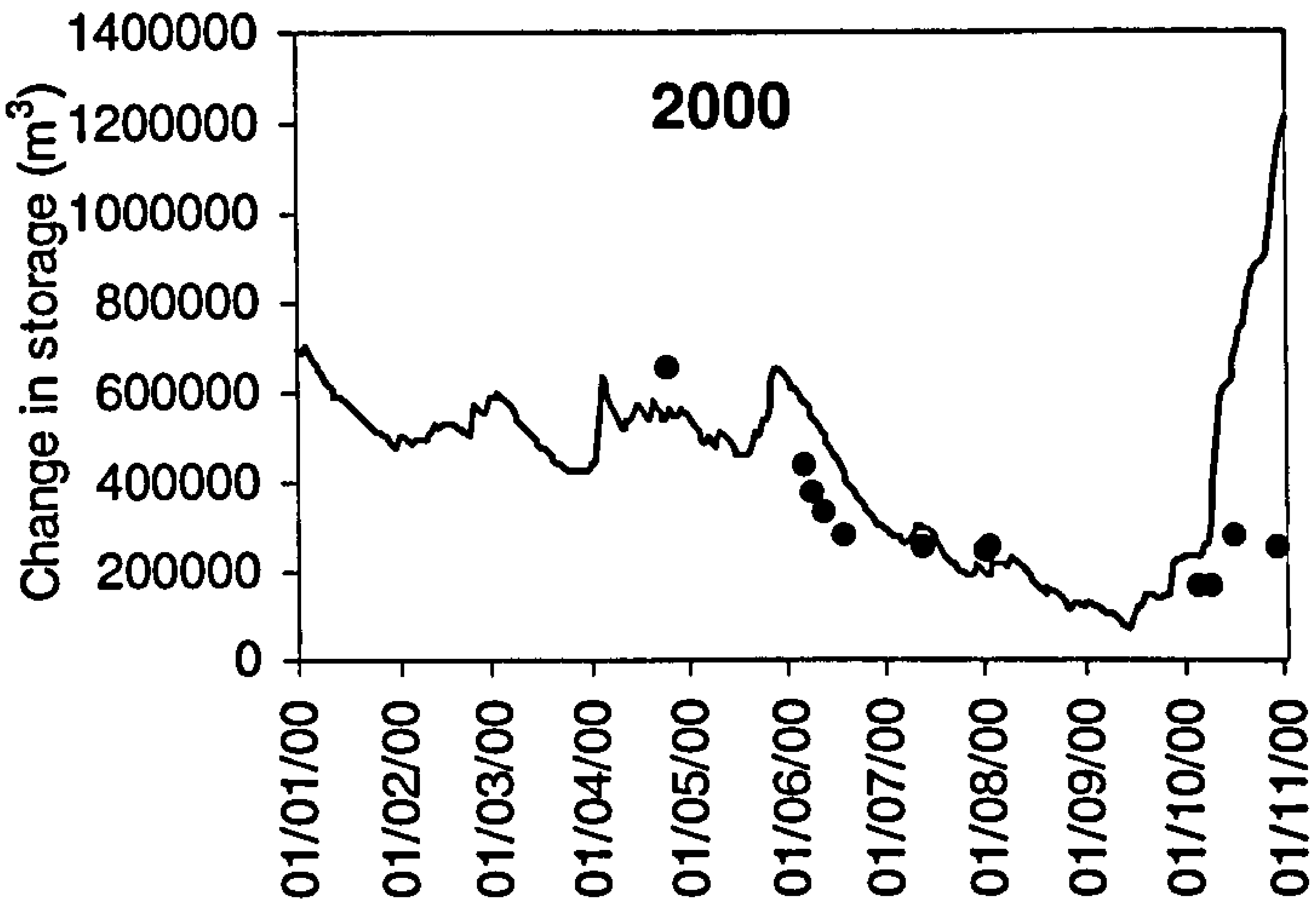

Figure I.04 – Modelled change in storage for 2000 compared with measured storage values (•).

I.3.1 Management implications

In general out of the 28 years studied (1974-2001) 19 (or two thirds) end up with a surplus of water in December and 9 result in a deficit, and these are the years that may require additional water from another source. The surplus or deficit of water on the site is plotted as a probability in Figure I.05.

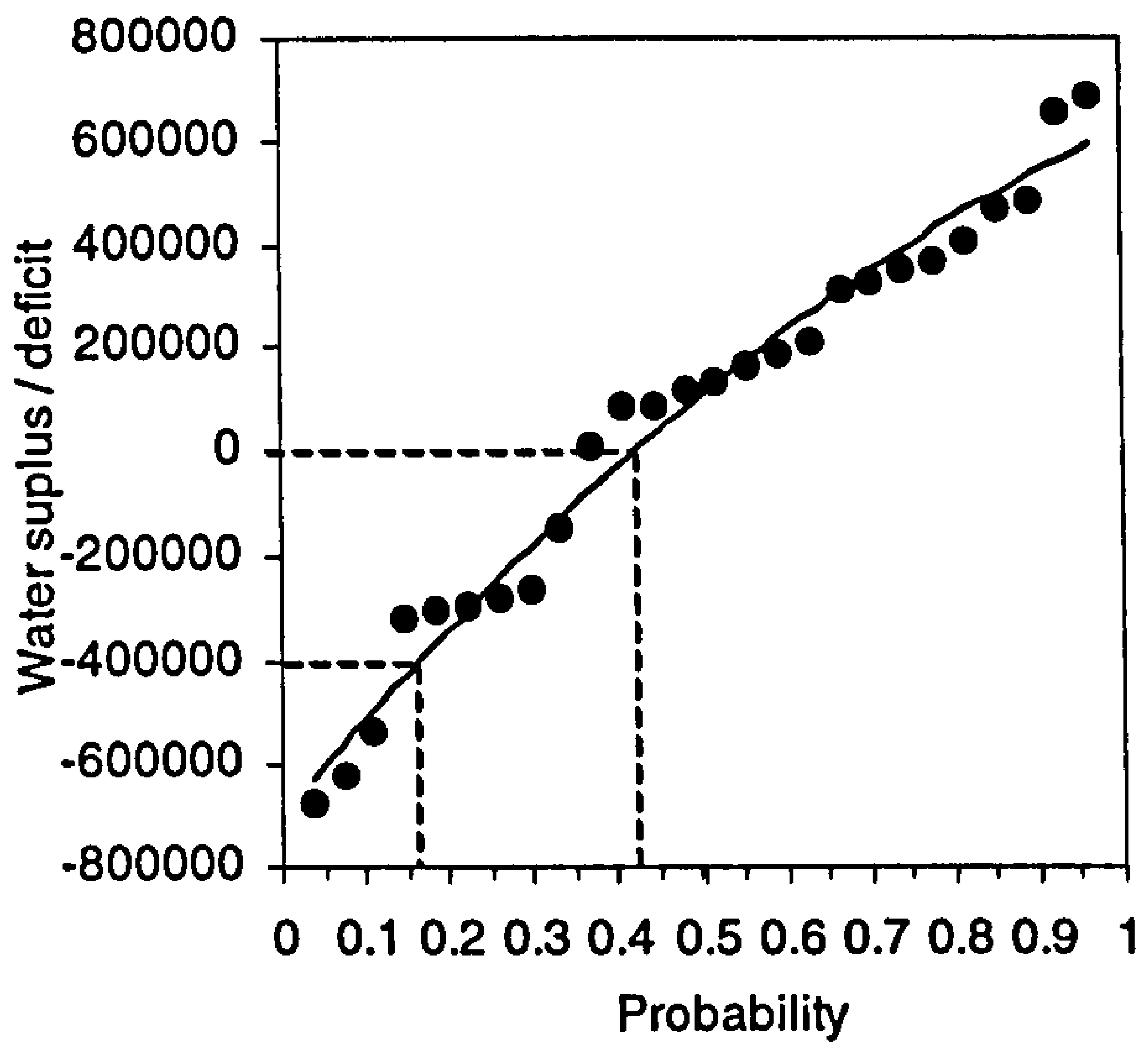

Figure I.05 – The probability of various water surpluses and deficits on Stodmarsh National Nature Reserve.

From Figure I.05 it can be concluded that if no extra water is added the site will be dry in around 42% of years. Although there are a number of factors that affect the amount of water that will be stored on the site it is possible to use the regression line of rainfall against annual storage change (Figure I.06) in order to calculate the change in storage that is likely for a particular annual rainfall amount.

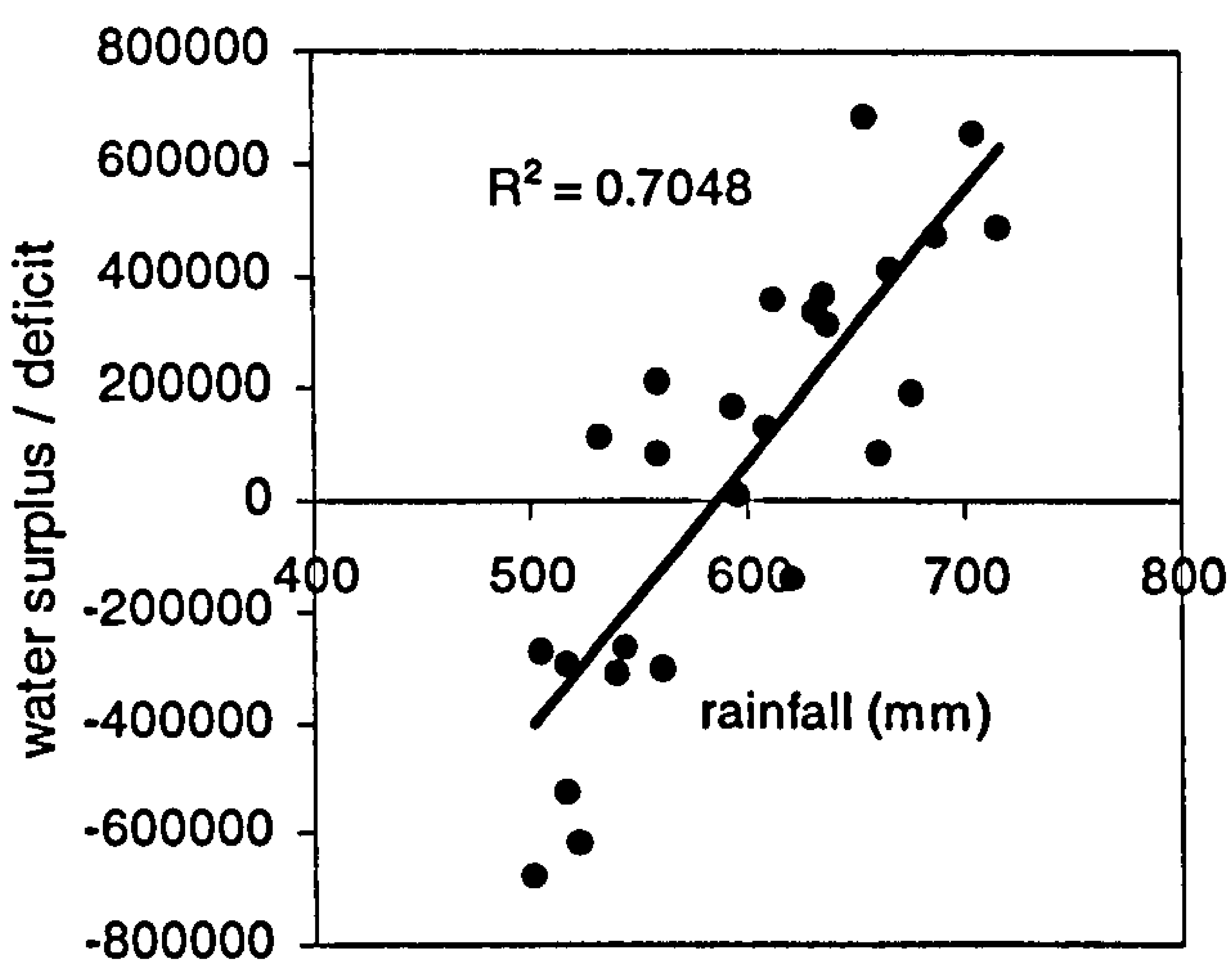

Figure I.06 – The relationship between total annual rainfall and change in storage.

The regression line is:

y = *4809.97 x – 2822090.85*

From this it is possible to work out the amount of water that will be required for different return periods of drought.

Table I.01 – The amount of water required in droughts of different return periods.

Drought recurrence (years)	Maximum water required (m^3)
1 in 3	0
1 in 4	200637
1 in 5	258857
1 in 6	319777
1 in 8	332005
1 in 9	332045
1 in 10	388726
1 in 38	530000

I.4 DISCUSSION OF THE WATER BALANCE MODEL

Despite the uncertainties within this approach, a model has been created that seems reasonable. Most years have the expected sigmoidal pattern in their storage volumes, with increases in the first few months of the year, a decrease due to evapotranspiration in summer and increases in the autumn and winter. Using the WaSim / Model Maker model there appears to be the sort of relationship that would be expected between total rainfall and change in storage (storage is greater in years of higher total annual rainfall) and which appears to fit fairly well into the measured data of 2000.

The imperfect relationship between rainfall and storage volume implies than total annual rainfall is not the only parameter that affects the amount of water stored on the site. Temporal rainfall distribution has also been shown to be important. For example, a year where the majority of the rain falls in the autumn and early winter will experience a much greater loss of storage than one with rain spread more evenly throughout the year. A reasonable amount of rain in spring seems to be important to maintain storage levels as it buffers against summer loss through evapotranspiration. As would be expected, years with high evapotranspiration loss have a greater loss of storage than those with a smaller loss.

From the this type of hydrological model it is possible to complete the primary aim of the thesis which was to calculate the amount of water used by Stodmarsh NNR in a year of particular rainfall. This can be done crudely using the regression equation of the line in Figure I.06. This predicts that an annual rainfall of greater than 587 mm will result in a surplus of water and less than this will result in a deficit. However in itself, this information is not that useful as it is impossible to predict what the rainfall will be in any given year and so apply the correct amount of water. Its usefulness is in risk prediction. The probability model shows that 42% of years will require some additional water and 16% of years will require greater than 400 000m^3 of water. In order to plan mitigation it is necessary to decide what return period of drought you are planning for – does the reserve need optimum water levels every year in order to avoid ecological disaster or can it cope with drought one in ten years, or even one in three years?

This is a pessimistic view because in the model each year is considered in isolation. It does not take account of the fact that in reality there is year to year continuity. The surpluses that occur in wet years have not been considered in the long term water budget and for example the impact of a dry spring may be lessened if it follows a wet winter and water levels are already high.

Possibly significant ecologically is the ability to predict maximum and minimum water table positions as this may indicate the potential for the wetland to be colonised by species preferring a more terrestrial habitat (Bradley 2002). *Phragmites* is a hardy species which can live in a wide variety of water regimes from water high above the surface to up to a meter below the surface. However if the reedbed is allowed to dry out other species will invade leading to a loss of the unique reedbed habitat and the consequential species such as the birds. For the site to have a mean water table at the surface at the end of the year this requires a storage loss of 395 927 m^3 of water. This occurs less than one year in ten, which indicates that invasion of other species into the ecosystem appears unlikely. Although in a number of years the mean water level does drop down to the surface level and below at the end of the summer, this is never for a duration long enough to change the character of the ecosystem. However if drought mitigation did not occur and there was a number of dry years consecutively – for example 6 of the 14 years with a water deficit have occurred since 1990 - the deficit of each would build on the last which could lead to longer term problems and possible ecosystem change.

I.5 ERROR ANALYSIS

The analysis below attempts to quantify the errors present within the annual results of this modelling approach. The majority of the analysis of the modelled water balance is done on an annual basis and therefore the errors are calculated over this time period. However none of the validation data is for a period as long as a year, so it must be assumed that the validation periods that do exist are representative of the whole year. Errors are calculated as the difference between the modelled and measured sum of the entire validation period. This is probably not unreasonable in the case of streamflow. However in winter evapotranspiration is effectively uncalibrated and unvalidated as no measurements were made in winter and therefore the errors are necessarily unquantified. However, since the magnitude of winter evapotranspiration is small, the absolute errors will also be small.

The biggest uncertainty concerning the annual water balance is input from the river. This has been estimated based on the measured water balances of 2000 and 2001, but as these were unusually wet years, this may be inappropriate in dryer years. Therefore errors may be up to 100%.

I.5.1 Rainfall

Errors in rainfall measurement were estimated as up to 16% on a daily basis. On a monthly and seasonal basis it would be expected that these errors would cancel out to some extent. A Monte Carlo analysis was carried out on 28 years of data, assigning each day a random error between 16% and –16% (uniformly distributed). The resulting annual values were summed and the percentage difference found as compared to the original annual sum. The mean value for all 28 months was found. This was repeated with 100 sets of random numbers and the mean error was 0.0097% (S.D. 0.00754).

I.5.2 Evapotranspiration

The errors in the evapotranspiration model were quantified by comparing the results of the model created using Penman Monteith reference evapotranspiration using data from Manston airport and crop coefficients from Fermor *et al.* (2001), with the seasonal average of the measured Bowen ratio evapotranspiration data collected in 2001 and 2002. The results are shown in Table I.02.

Table I.02 –Seasonal percentage error in the validation of the evapotranspiration model

Year	Month	Modelled ET (mm)	Measured ET (mm)	% error
2001	Seasonal	2.03	1.96	3.88
2002	Seasonal	2.01	2.18	7.78
Mean seasonal error				**5.83**
S.D.				8.80

I.5.3 IHACRES inflow model

The IHACRES inflow model unfortunately only has a very short validation period – there are only 3 months of complete data, therefore it is difficult to be confident of the error estimation. Over the whole period the error is 7.8%.

I.5.4 WaSim / ModelMaker outflow model

Due to the difficulties caused by flooding, the outflow model is probably the least certain part of the hydrological model. However it does have a long validation period with which to estimate the magnitude of the problem. The error over the entire validation period is 19.6%.

I.5.5 Water balance errors

The water balance is an additive model and therefore the total errors are calculated as the sum of the absolute errors in each component. Figure I.07 summarises the annual relative errors.

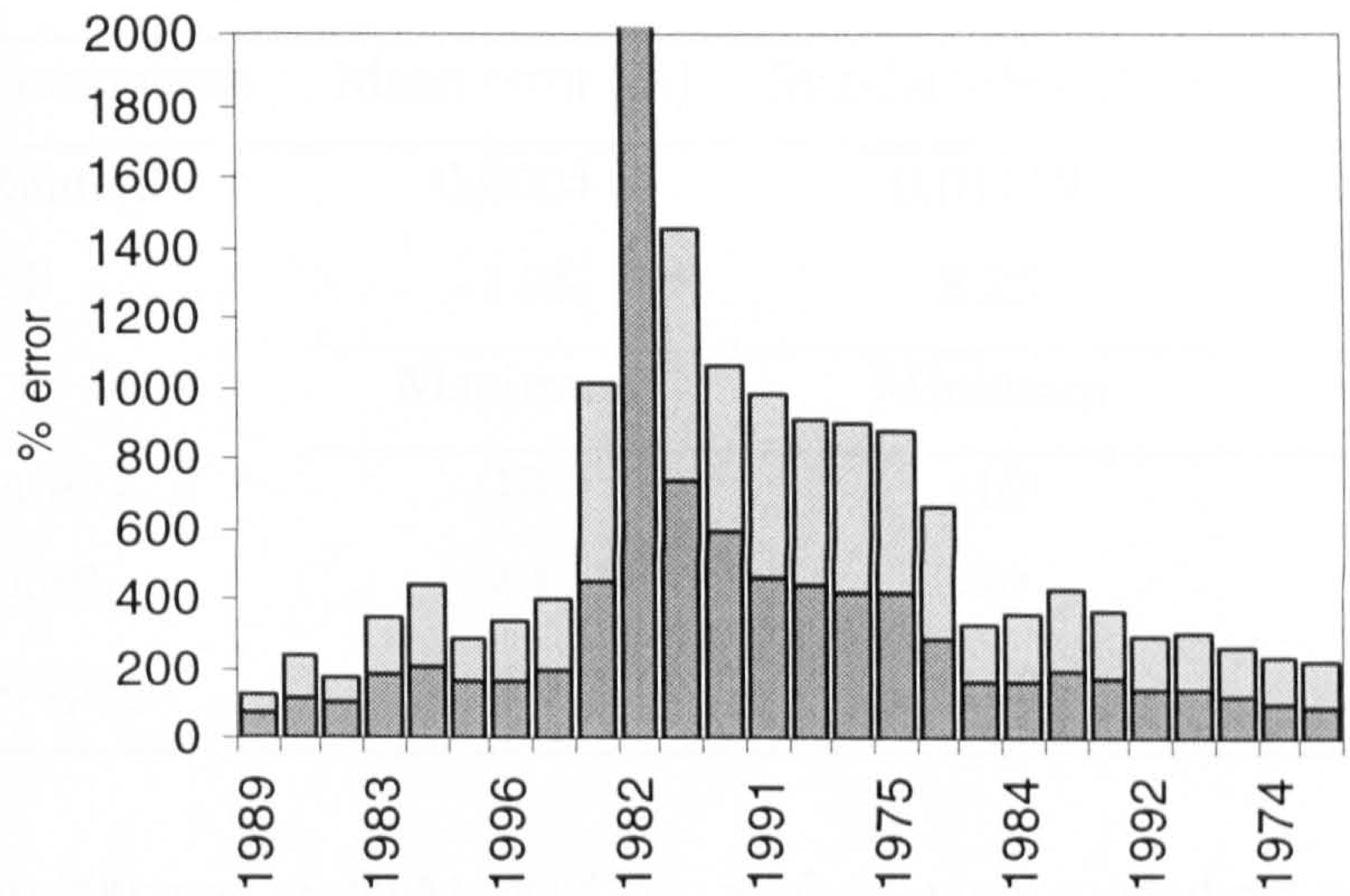

Figure I.07 – Annual water balance errors arranged in order of volume of storage change (increasing left to right). The entire bars represent the relative errors including the errors caused by seepage. The darker parts of the bars are the relative errors when seepage is ignored.

It can be seen that the errors are consistently large. Seepage represents about one third of the errors. Again there is a relationship with storage change volume, with those that change very little in a year having the highest uncertainty.

I.5.5.1 Most likely annual errors

The above analysis does not account for any cancelling out of errors between components. Therefore a Monte Carlo analysis was carried out to determine the most likely range of errors. In the case of rainfall this was carried out using the mean and standard deviation of errors worked out in the Monte Carlo analysis described in section 8.5.1, assuming a normal distribution of errors. For evapotranspiration the mean and standard deviation of the annual results for 2001 and 2002 was used. For streamflow a uniform distribution was used as there is only one season of data and therefore we have no real idea of the range of errors that may occur over a long period. Table I.03 shows the values used.

Table I.03 – Parameters used to specify the range of errors in each component of the water balance

Component	Mean error (%)	Standard deviation	Distribution
Rainfall	0.0004	0.01159	Normal
ET	-1.95	8.25	Normal
	Maximum	Minimum	
Inflow	10	-10	Uniform
Outflow	40	-40	Uniform
Seepage	100	-100	Uniform

50 000 runs of the Monte Carlo analysis were carried out using Crystal Ball.

The frequency chart of the mean error of all the years studied is in Figure I.08.

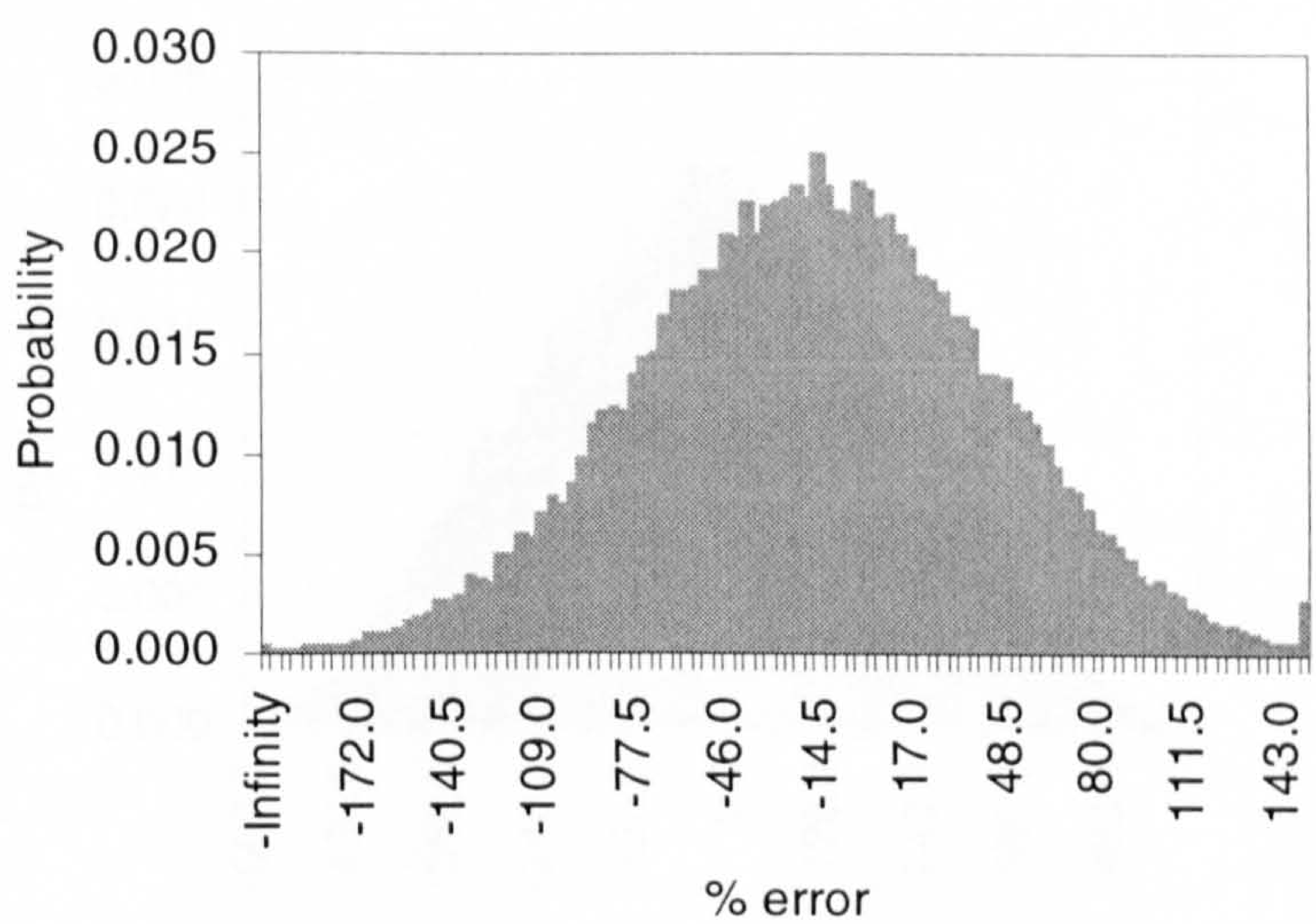

Figure I.08 – Frequency histogram of the mean percentage error of all the years studied.

The mean error is –13.9% (S.D. 59.3). It is possible to be 95% certain that the mean error of all years will lie between –130.8% and 101.9%.

The Monte Carlo analysis was also run for individual years to investigate the potential range of errors. The results of this are presented for three years – a year with a large change in storage (1989), a medium change in storage (1994) and a small change in storage (1980). The three frequency charts are shown (Figure I.09-I.11) and the statistics are shown in Table I.04.

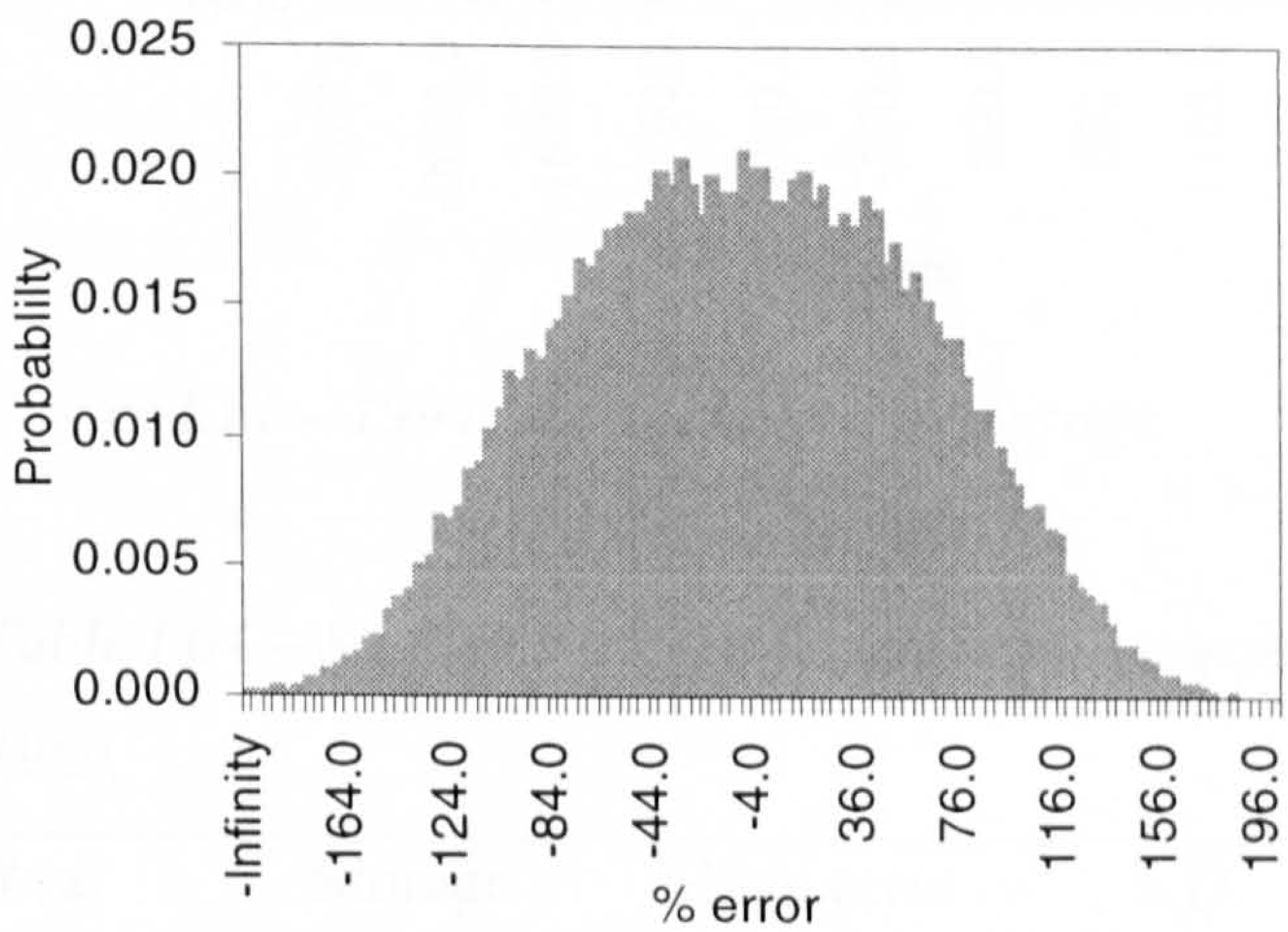

Figure I.09 – 1989, a year of large change in storage

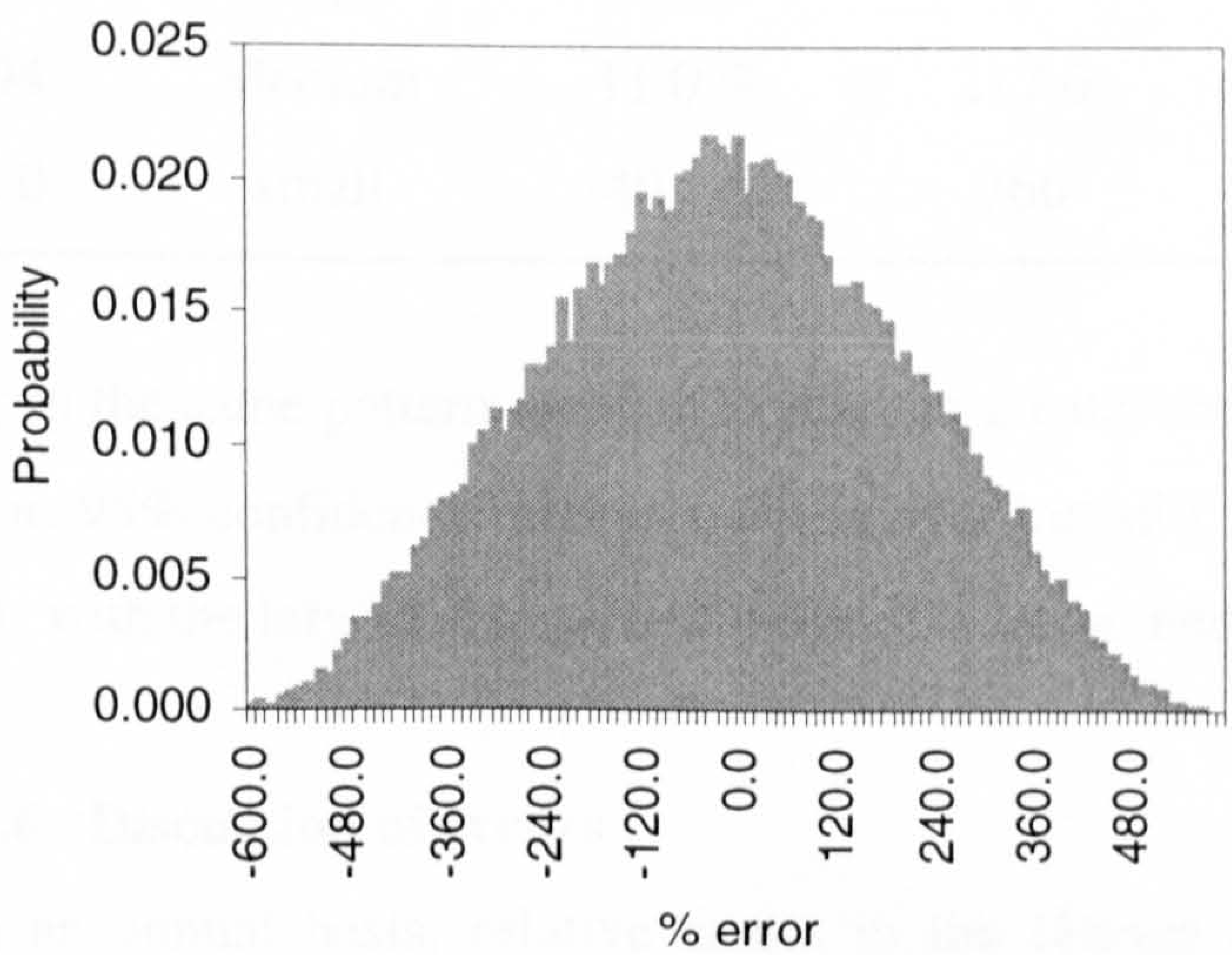

Figure I.10 – 1994, medium change in storage

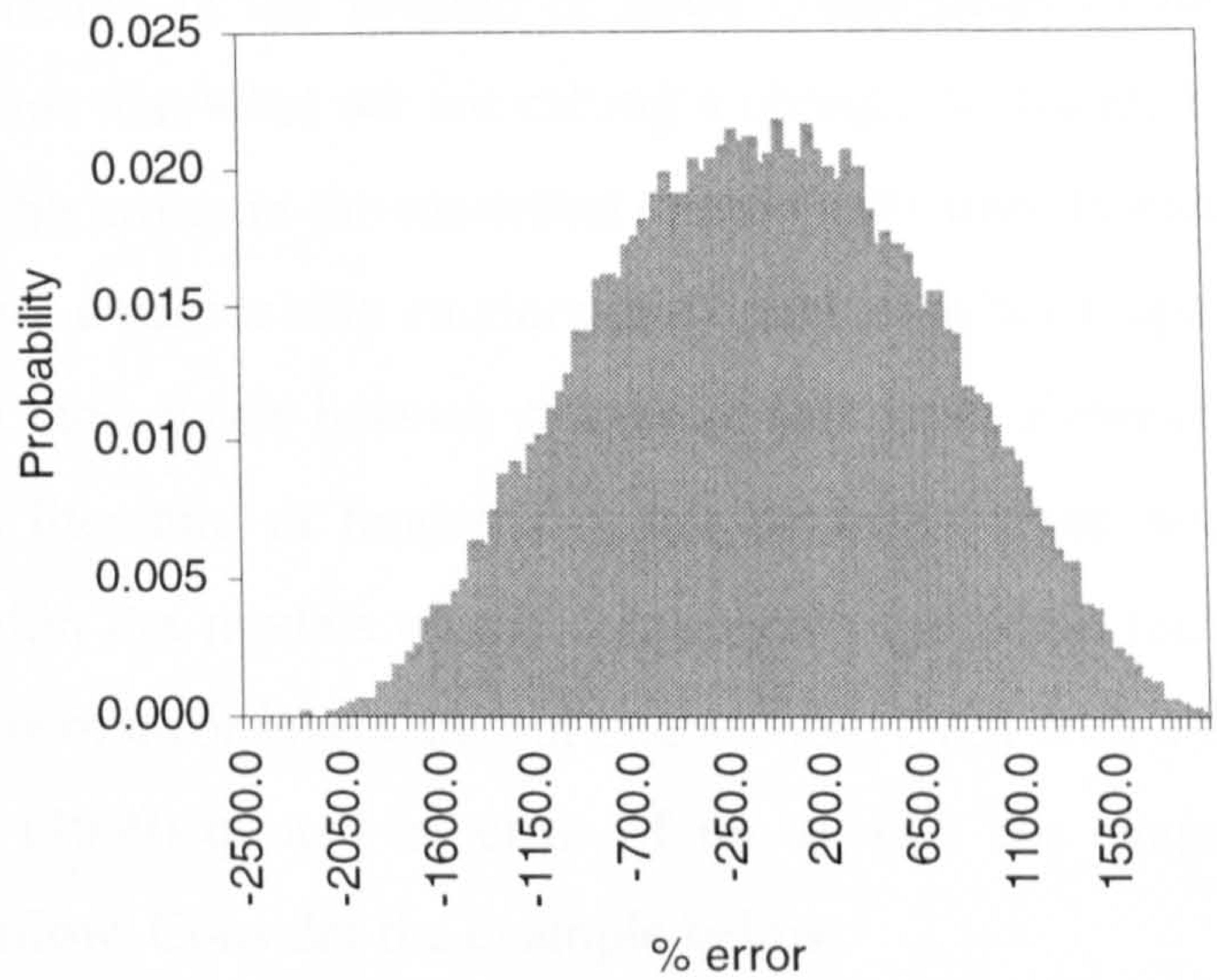

Figure I.11 – 1994, small change in storage

Table I.04 – Statistics for the frequency histograms of the errors for 1989, 1994 and 1980.

Year	Storage change	Mean error	S.D.	95% confidence Min	max
1989	Large	-5.8 %	69.1	-134.0 %	123.3 %
1994	Medium	11.0 %	217.6	-424.6 %	401.9 %
1980	small	-40.1 %	766	-1504.7 %	1394.3 %

Again the same pattern appears in that the greater the change in storage, the smaller the error. 95% confidence intervals are smaller but still show a large margin for error. The year with the largest change in storage also has a mean error closest to 0.

I.5.6 Discussion of errors

On an annual basis, relative errors in the change in storage volume are very high, particularly when the change in storage is small. In every case the 95% confidence intervals include greater than 100% error. This is because as the volume of change in storage is very small compared to the sum of the volume of the other components, and therefore the sum of their absolute errors are in almost every case bigger than the change in storage itself. Absolute errors can remain large in comparison to the volume of the change in storage, as a small change in storage is not necessarily related to reduction in the volume of other components of the water balance. This effectively means that what we are calling a change in storage could in fact be entirely accounted for by errors in the modelled components used to calculate that change in storage. This is not a particularly encouraging conclusion but it appears that it may be true of many, if not most water balance studies of this type. However no error analyses were found in the literature of modelled water balances. Even with substantially lower error levels within the models of the components the same problem will arise. In addition, these sorts of error levels are not uncommon. Even with a measured water balance, Drexler *et al.* (1999) quoted an error of up to 40% for evapotranspiration and 44% for steam outflow. Consider the example below:

Over the course of a year if outflow was overestimated by 10% and ET also overestimated by 10% this would result in an additional 421349.7 m^3 of water in the model outputs compared to reality (based on 1990 data), the equivalent of 173.5 mm of water and greater than the total storage change in 19 of the 28 years studied. If the errors were only 5% the change would be 86.7 mm of water, greater than the storage change for 8 of the years studied.

Errors have been found (and assumed) to be random rather than systematic and it is very important to ensure that this is the case and that any systematic errors are removed (this was considered in the model calibration in Chapter 6 when using goodness of fit statistics). Also the error assigned to the seepage estimation of 100% every year is also almost certainly an over-estimation, as it is known from on site research, and from other people's long term experience of the site that seepage is an issue. Seepage accounts for one third to one half of annual errors when maximum errors are found.

Appendix J – Description of Data on Attached CD

The CD attached to this thesis contains raw data collected during fieldwork in the course of the research. This appendix describes the contents of the CD. It is summarised in Table J.01 below for easy reference and more detail is given on some of the files below.

Table J.01 – List of files contained on the raw data CD

File name	Data contained within file
Bowen ratio data.xls	Meteorological measurements taken using the Bowen ratio equipment in 2001 and 2002.
Height of reeds.xls	Measurements of the height of individual reed stems taken on dates in 2001 and 2002
IRGA results.xls	Results produced by the IRGA for each individual measurement taken on four separate sampling days
Kent inflow raw data.xls	Discharge data produced by the Unidata Starflow
Kent outflow raw data.xls	Discharge data produced by the Unidata Starflow
LAI values.xls	Individual measurements made using the sunscan on three separate days
Windspeed data.xls	Data measured using three anemometers at various heights and calibration data from the anemometers that need it.
Water depth measurements sub-folder	Shapefiles and DBF files of water depth measurements taken using GIS equipment, which can be plotted on Basemaps in ArcView

Bowen ratio data.xls

The workbook contains two worksheets, one for data collected in 2001 and one for data collected in 2002. Each sheet contains the same columns of data. The data is described in Table J.02.

Table J.02 – Description of the raw data given in the Bowen ratio data.xls file. The letters refer to the column labels in Microsoft Excel.

	Data		Data
A	Datalogger code	G	Change in soil temperature (°C)
B & C	Julian day and time of reading	H	Datalogger code
D	Net radiation (W m^{-2})	I & J	Temperature at lower and upper (high) thermocouples (°C)
E	Ground heat flux measured by soil heat flux plate (Wm-2)	K & M	Dewpoint temperature at lower and upper (high) arms (°C)
F	Soil temperature (°C)	L & N	Vapour pressure at lower and upper (high) arms (kPa)

IRGA results.xls

Each worksheet contains the data for one day's measurement. Each worksheet contains data as described in Table J.03.

Table J.03 – Description of the raw data given in the IRGA results.xls file. The letters refer to the column labels in Microsoft Excel.

	Data		Data
A	Level in the canopy at which measurement is taken	J	Cuvette air temperature (°C)
B	Leaf number within sample group	K	Leaf area within cuvette (cm^2)
C	Date	L	Flow rate of air (ml min^{-1})
D	Time	M	Transpiration (mmoles m^{-2} s^{-1})
E	Reference CO_2 concentration (ppm)	N	Stomatal conductance (mmoles m^{-2} s^{-1})
F	CO2 differential (ppm)	O	Leaf temperature (°C)
G	Photosynthetically active radiation (mmol m^2 s^2)	P	Calculated CO_2 uptake (ppm)
H	Reference water vapour pressure (mb)	Q	Calculated sub stomatal CO_2 concentration (ppm)
I	Differential water vapour (mb)		

Kent inflow raw data.xls and Kent outflow raw data.xls

The inflow and outflow raw data are presented in separate files due to the large quantity of data collected. Readings were taken every twenty minutes over the course of around two years. The data included are described in table J.04

Table J.04 – Description of the raw data given in the Kent inflow raw data.xls and Kent outflow raw data.xls file. The letters refer to the column labels in Microsoft Excel

	Data		Data
A	Date	E	Mean depth (mm)
B	Time	F	Mean temperature (°C)
C	Maximum velocity (mm s^{-1})	G	Mean velocity (mm s^{-1})
D	Minimum velocity (mm s^{-1})	H	Number of samples

Windspeed data.xls

Data is given for 2001 and 2002. After the datalogger code and the date and time of measurement, three columns of data are shown, one for each of the anemometers in the wind profile. The height at which the measurements were taken are given at the top of the columns. The first column of data represents the lower anemometer and this measures in m s^{-1}. However the other anemometers just display a count of rotations and this must be converted in to m s^{-1} using a calibration equation. These are given for each year in a separate worksheet.

Water depth measurements sub-folder

This folder contains shapefiles and dbf files created whilst taking water depth measurements in combination with GIS data. This data is presented in a form which can be plotted as a point theme onto a basemap in ArcView GIS (Environmental Systems Research Institute Inc. version 3.2a).

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